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GEOLOGICAL MAGAZINE

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GEOLOGICAL MAGAZINE

VOL. XCVIII, No. 1. PUBLISHED 3.2.1961 JANUARY–FEBRUARY 1961

Ultramafic Breccias from the Musa Valley, Eastern Papua

By D. H. GREEN*

(PLATES I and II)

ABSTRACT

In the Musa Valley area rocks of the Papuan Ultramafic Belt outcrop discontinuously over an area of 25 miles by 45 miles. The rocks are part of a folded layered sequence ranging from magnesian dunite and peridotite, chemically and petrographically typical of the alpine ultramafic suite, to olivine gabbro and bytownite gabbro. Agglomerate-like breccias consisting of fragments of ultramafic rock in a variable matrix occur as irregular vent-like bodies in the peridotite and dunite and as horizontal sheets in a Pleistocene-Recent sedimentary sequence. The breccias are interpreted as vent and extrusive breccias resulting from the penetration, brecciation and local entrainment (fluidization) of peridotitic country rock by volcanic gases. Olivine alkali basalt was probably the parental magma responsible for the gaseous activity. The breccias show a sequence of stability fields for the assemblages mesh-texture serpentine and bastite (i.e. chrysotile and lizardite mixtures); antigorite; olivine and enstatite; and these are correlated qualitatively with variation in intensity of the brecciation process. Chemically the brecciation process was responsible for the addition of H_2O, SiO_2, CO_2, Al_2O_3, and O_2 and the removal of Fe and Mg from the fragmented rock. Minerals which are of direct crystallization in the matrix fluid environment include chalcedony, fibrous serpentine (chrysotile), magnesite, quartz, andradite, and garnierite. There is no evidence of the presence of an ultramafic, serpentine magma in the genesis of the breccias.

INTRODUCTION

THE presence of ultramafic breccia, tentatively regarded as a fault breccia, in the Musa Valley area of Eastern Papua, was noted by J. Thompson of the Australian Bureau of Mineral Resources during a reconnaissance traverse in the area during 1957. In 1958 a B.M.R. field party consisting of J. W. Smith and the writer spent five months in the geological mapping of the Musa Valley area, (Text-figs. 1 and 2). Much of the area is strongly dissected and mountainous terrain with tropical vegetation. The Goropu Mountains in the south-east rise sharply from 2,000 to 11,000 ft.; mountains in the south-west and north-west are up to 7,000 ft. and the Didana Range north of Wowo Gap (about 1,500 ft.) is 4,000–5,000 ft. Outcrops and geological traverses were largely limited to accessible streams and a small number of tracks between native villages.

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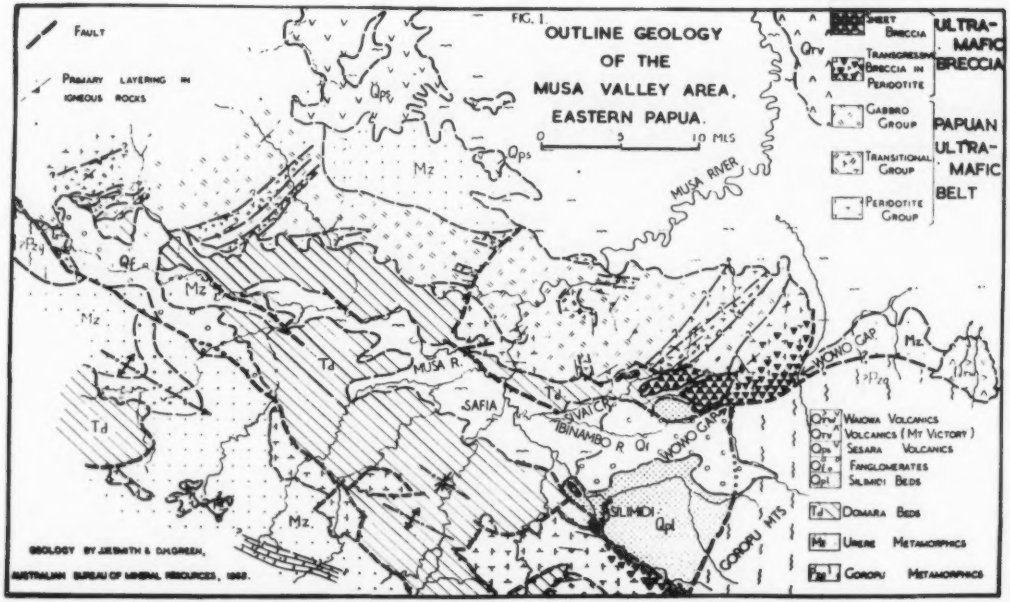
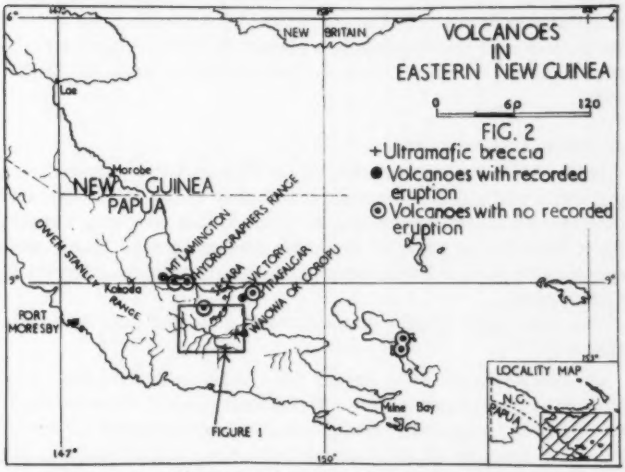


TEXT-FIG. 1.—Geological map of the Musa Valley, Papua.

General Geology

South-eastern Papua forms part of the circum-Pacific Tertiary and Recent orogenic belt and its sedimentation, volcanism and occurrences of ultramafic rocks are characteristic of this environment. The Papuan Ultramafic Belt lies on the northern side of a belt of ?Palaeozoic metamorphics forming the main range of Eastern Papua (Owen Stanley Range). The outcrops in the Musa Valley mark the eastern extremity of the main part of the Ultramafic Belt but ultramafic rocks, probably of the same " serpentine belt ", occur near Milne Bay at the eastern tip



TEXT-FIG. 2.—Locality map and location of volcanoes in Eastern Papua.

of Papua. The Papuan occurrences would lie on the New Zealand-New Caledonia serpentine belt of Hess (1955).

Table 1 summarizes the stratigraphic and igneous sequence at the eastern end of the Musa Valley. It is to be noted that the sheets of ultramafic breccia [Sivai Breccia] within the Silimidi Beds considerably post-date the emplacement and exposure to erosion of the ultramafic-basic suite and are separated in time therefrom by two periods of sedimentation, interrupted by major faulting.

1. *Volcanic Activity.*

There have been three phases of volcanic activity in the area since the emplacement of the ultramafic-basic suite. During the deposition of the Domara River Beds, hornblende andesitic lavas, tuffs and agglomerates were deposited near several local centres and this

activity continued to a final phase of dyke intrusion during the post-depositional faulting.

At Wowo Gap there is evidence of olivine basaltic activity in association with the ultramafic breccias and 30 miles to the north-west, on the northern flank of the ultramafic belt, a Pleistocene-Recent volcanic cone is composed of extrusions and fragmental deposits of a petrographically identical olivine basalt (Sesara Volcanics).

The Waiowa Volcanics occur near a small active vent (Goropu Volcano of Baker (1946) and Taylor (1958)) about 12 miles east of Wowo Gap. These volcanics are petrographically similar to a series of small lamprophyre and biotite pyroxenite dykes intruding the Domara River Beds and the hornblende andesitic intrusives. They are also similar to the Mount Lamington (1951) eruptive material (Taylor, 1958).

2. *Papuan Ultramafic Belt*

In the Musa Valley area rocks of the Papuan Ultramafic Belt crop out over a width of 25 miles and a length of 45 miles. Within this area outcrops are discontinuous due to cover of the overlying Domara River Beds and to areas of the Urere Metamorphics. Where seen, contacts with the Urere Metamorphics are sheared contacts of uncertain displacement.

The rock types of the ultramafic-basic suite fall into four main groups:

(*a*) *The peridotite group* ranges from massive dunites and harzburgites to layered peridotites with rhythmic layers 3–12 inches thick. The minerals, magnesian olivine, enstatite and accessory red-brown chromite, typically show undulose extinction, strain banding and lamellae, and intercrystal granulation to a varying degree.

(*b*) *The transitional group* is a strongly layered and very varied sequence including peridotite, pyroxenite, feldspathic peridotite, norite and gabbro. The minerals present include olivine, enstatite, hypersthene, diallage, bytownite, magnetite, chromite and primary amphibole. The compositional banding varies in thickness from over 50 feet down to less than 1 inch (bands of 2 or 3 crystals in thickness). Textural evidence of crystal accumulation is provided in some rocks by euhedral and sub-hedral olivine and orthopyroxene with interstitial, large anhedral plates of bytownite and diallage or primary amphibole.

(*c*) *The gabbroic group* consists largely of massive medium-grained bytownite (An_{80}-An_{72}) gabbro and olivine gabbro but also includes rhythmically banded gabbro showing typical gravitational layering with basal accumulative ferromagnesian minerals in the layers.

(*d*) *The pegmatitic pyroxenite and gabbro group* occurs in several

TABLE 1.—STRATIGRAPHICAL TABLE, EASTERN MUSA VALLEY, PAPUA

Unit	*Age*	*Description*
Waiowa Volcanics	1943–44	Local agglomerate and tuff of augite biotite andesite.
Ubo Fanglomerate	Recent	Unconsolidated fanglomerate deposits, largely of derived Goropu Metamorphics, overlying in part the depositional surface of the Silimidi Beds and in part an erosional surface.
		Local Faulting
Silimidi Beds	? Pleistocene—Recent	Interbedded polymictic conglomerate, sedimentary breccia, and greywacke. Contains Sivai Breccia member consisting of at least two sheets of ultramafic breccia with interbedded conglomerate and sedimentary breccia.
Unconformity at Sivai Cr. ? Disconformity near Silimidi		Major faulting established present topographical outline. Common intrusion of hornblende-feldspar porphyry along faults.
Domara River Beds	? Pliocene—Pleistocene	Interbedded polymictic conglomerate, greywacke, and siltstone, Several local, usually basal, volcanic members.
Major Unconformity		2. Ultramafic-mafic belt emplaced at present crustal level and exposed to rapid erosion into the Domara River Beds. 1. Granodiorite-diorite intrusion into Urere Metamorphics; low grade metamorphism.
Urere Metamorphics	? Cretaceous	Low grade, usually non-schistose metamorphics—largely dolerites, basalts, pillow lavas; some greywacke and siltstone; Foasi Limestone and Amora Conglomerate members.
Possible Unconformity		Fault contact.
Goropu Metamorphics	?	Quartz-mica schists and phyllites, chlorite schists, quartzites; metamorphosed (talc, antigorite) older serpentinite.

irregular bodies consisting of patchily distributed, very coarse, pyroxenite, pegmatitic gabbro and some serpentinite.

The first three groups form part of a related sequence of rocks and their contacts are interlayered and transitional, and concordant in the attitude of their layering. The sequence of rocks does not seem to be a single, simple sequence of ultramafic to basic rocks for there are apparent repetitions of the groups and the transitional group may occur with mainly gabbroic rocks both above and below.

The fourth group appears to be intrusive into the layered sequence. Its pegmatitic character and variation through to a hornblende gabbro in which final crystallization was of oligoclase-andesine rims and colourless amphibole suggest that this may represent a more volatile-rich and possibly residual magma.

The nature of the layered series, its banding and textural features, indicate an accumulative, gravitational differentiation origin from a magma or series of magmas of uncertain composition. The compositional layering is folded, dipping at shallow to steep angles and variable in strike. No simple structural pattern could be defined although the strike of the layering is generally inclined to the strike of the ultramafic-basic belt as a whole. These features, coupled with the strain effects very evident in thin sections of the dunites and peridotites and to a lesser extent in the gabbros, show that there has been major deformation of the rocks following their initial consolidation.

Serpentinization of the ultramafic rocks is of irregular distribution. In the Wowo Gap area there are some completely serpentinized outcrops and south of the Musa Valley bodies of ultramafics which are fault bounded within the Domara River Beds or in sheared contact with the Urere Metamorphics, are commonly extensively serpentinized.

Antigorite serpentinite has only been observed in two environments—within the ultramafic breccias and in several metamorphosed, presumably much older, serpentine-talc bodies within the Goropu Metamorphics.

Ultramafic Breccias

1. *Field Occurrence.*—Agglomerate-like breccias consisting of angular fragments of ultramafic rocks in a variable matrix have two modes of occurrence in the area.

(*a*) *Transgressive breccias.*—These bodies of breccia occur within peridotite showing a variable degree of serpentinization. The bodies are typically irregular and non-linear in outcrop; more rarely they have linear margins controlled by intersecting joint planes in the peridotite. Usually there is a transitional passage from marginal massive peridotite, through finely shatter-jointed peridotite and matrix-poor breccias to

central breccias in which fine (<2 mm.) matrix material may comprise more than 40 per cent of the rock. These transgressive bodies or pipes show an extreme variation in the size of neighbouring fragments and contain blocks over 4 ft. in diameter. Initial stages of further fragmentation are often evident in these larger blocks. A striking feature about the field occurrence is the lack of shearing and slickensides either marginally or within the breccia bodies. There are compositional and textural variations in the breccia bodies which are apparently related to the intensity of the brecciation process.

The transgressive breccias are most abundant at and north of Wowo Gap and in a second area south of Silimidi. In both these areas a strong N.N.E. fault zone (probably transcurrent) intersects normal faults trending E.–W. to E.S.E.–W.N.W.

(*b*) *Sheet Breccias.*—Conformable sheets of hard, agglomerate-like ultramafic breccia occur within the Silimidi Beds at the eastern and south-eastern edges of the Musa Valley. In these two areas there are at least two sheets of the breccia of approximate thickness 70 ft. and 30 ft. separated by poorly consolidated polymictic conglomerate, sedimentary breccia and greywacke or tuffaceous sandstone. The two areas are separated by alluvial and fanglomerate deposits of the Ibinambo River and may or may not originally have been continuous. The depositional surface of the Silimidi Beds is largely preserved as a topographical feature and slopes upwards to Wowo Gap at about 3°–5°. At Wowo Gap the sheet breccias approach the same topographic level at which the transgressive breccias occur. However, south of Silimidi the Silimidi Beds, the interbedded sheet breccias, and the underlying Domara River Beds are in normal fault contact with peridotite and dunite containing bodies of transgressive breccia.

2. *Petrography.*—The ultramafic breccias may be grouped into several types based on the nature and degree of alteration of the fragments but a number of observations apply to all the types.

In none of the thin sections of the breccias examined is there any evidence of shearing and schistosity either in the matrix or on the margins of fragments. This is applicable whether the fragments are of unserpentinized peridotite and dunite, or fragments and matrix are of serpentine, a mineral readily showing the effects of shearing. Contrasting with this, the dunite and peridotite fragments internally show the strong and variable development of stress effects as described previously in the peridotite group of rocks. These effects pre-date any serpentinization present and in some cases incipient mylonitic texture has localized serpentinization, so accentuating the mylonitic striping visible in hand specimens.

The fragments are characteristically angular but in some peridotite and antigorite breccias, particularly the sheet breccias, there is slight

smoothing and rounding of moderately large fragments. Fragments, whether large or very small are rarely in contact with one another but are separated by non-fragmental matrix material. Another feature commonly observed is that fragments of one fractured mineral grain may show only limited dispersion one from another although the grain is completely fragmented. This is particularly well shown by chromite which occurs as uncommon medium to coarse crystals in the peridotite but in the matrix to the breccias commonly occurs as clusters of small fragments of the original larger crystal. Enstatite also illustrates the same relationship and the observation is valid in the sheet as well as the transgressive breccias.

(*a*) *Nomenclature and Identification of Serpentine Minerals.*—In the thin section examination of the breccias a clear distinction (corresponding to that described and illustrated by Francis (1956)) could be made between serpentinites possessing a random thorn or flame texture typical of antigorite, and serpentinites possessing a pseudomorphic texture of serpentine in bastite form (after orthopyroxene) and, more abundantly, in mesh-texture form (after olivine). In many thin sections the serpentine in antigoritic texture clearly has a replacement relation to the mesh-texture and bastite serpentine mixture; in others the textural types are mutually exclusive.

X-ray powder diffraction data* on these two textural types indicates a mineralogical distinction, the powder patterns falling into two groups with no mixing of the contrasted patterns. (The specimens selected were those which in thin section were completely composed of one or other textural type.) The antigorite textural type has a pattern corresponding very well with antigorite patterns given by Whittaker and Zussman (1956). Similarly the mesh-texture and bastite serpentine mixtures gave patterns typical of mixtures of clinochrysotile and lizardite as defined by Whittaker and Zussman. Intensity distributions indicate that in some mixtures clinochrysotile is in excess of lizardite, in others the reverse is true. The diagnostic characters of the two groups are that antigorite has a strong line at d = 1·563 Å [24,3,0] forming a pair with a moderately strong line at d = 1·541 Å [0,6,0] and this pair can be distinguished from the moderately strong lizardite + clinochrysotile pair with d = 1·536 Å ((060) for lizardite, orthochrysotile, and clinochrysotile) and d = 1·502 Å ((208) for lizardite and orthochrysotile only). Confirmation is found in the presence of a strong antigorite line at d = 2·52 Å (16,0,1) which is distinct from the strong lizardite line at d = 2·50 Å (202). Confusion may result from the use of this criterion alone as any olivine impurity in the sample

* The powder photographs were taken with a 19 cm. camera using Ni-filtered radiation from a Cu target. Measurements were corrected for film shrinkage.

will give a strong line at d = 2·514 Å which may be confused with the antigorite line.

The absence of orthochrysotile in these mixtures is indicated by the absence of 20*l* reflections for "*l*" odd and their presence for "*l*" even (Whittaker and Zussman, 1956).

In this example the X-ray powder diffraction data given by Whittaker and Zussman for the identification of antigorite, clinochrysotile, orthochrysotile and lizardite led to unambiguous results when applied to rock-forming serpentine minerals. These results are consistent with the petrographic distinction between antigoritic and mesh-texture + bastite serpentine textures and the absence of mixtures of antigorite with chrysotile + lizardite in the selected specimens confirms the reaction and replacement relationship between the textural types as seen in thin section.

(*b*) *Mesh-Texture Serpentine Breccia.*—In this type of breccia fragments show partial and complete stages of replacement of olivine and orthopyroxene by mesh-texture and bastite serpentine respectively. Antigorite is absent. The mesh-texture serpentine shows colour and birefringence variation due apparently to the exsolution of magnetitie as distinct grains or as very finely divided dust. In the least brecciated sections the partially serpentinized peridotite is veined and fragmented by dilational, colloform chalcedony veins. More typical mesh-texture serpentine breccias show more advanced brecciation and accompanying serpentinization. Fragments of mesh-texture serpentine occur down to less than 0·01 mm. grain size but interstitially to these there is very, very fine, amorphous and fibrous, sub-isotropic material which is in part chalcedony and in part a finely granular or fibrous, yellow brown, isotropic or sub-isotropic material, probably a chrysotilic serpentine mineral. Vughs are quite common and usually lined with colloform silica; the colloform habit is also present encircling small fragments.

(*c*) *Antigorite Serpentine Breccia.*—The fragments in this breccia show partial and complete replacement of mesh-texture and bastite serpentine and of orthopyroxene and olivine by antigorite. In some specimens the antigorite directly replaces olivine and orthopyroxene; in others there has been prior alteration to mesh-texture and bastite serpentine. In both cases the antigorite typically has the characteristic random thorn or flame texture (cf. Francis, 1956) or more rarely shows a dominant growth orientation giving the "bowenite" form. In a number of fragments the antigorite occurs in radiating flares centred on the fragment margin or in laths perpendicular to the margin. In these cases antigoritization probably post-dates brecciation but in other cases antigorite fragments have been further brecciated following antigoritization.

Fragments occur down to less than 0·05 mm. grain size but

interstitially to these there is a variable proportion of turbid, yellow brown, sub-isotropic, very fine material consisting of at least two minerals, chalcedony and a fibrous serpentine mineral. Colloform chalcedony occurs locally as vugh linings and encircling fragments. The serpentine mineral is well crystallized and its form clearly shows that it is a direct crystallization in the matrix and is not fragmental. From its optical properties* and from its X-ray powder diffraction pattern the mineral is identified as clinochrysotile having the characteristic reflections 202 (d = 2·45 Å), 204 (d = 2·09 Å), 206 (d = 1·74 Å), 060 (d = 1·53 Å) and $40\bar{2}$ (d = 1·315 Å).

A small amount of magnesite is present as a matrix crystallization in some breccias and in one specimen there are a few sub-hedral and euhedral garnets associated with well-crystallized patches of the matrix clino-chrysotile. This garnet is andradite (colourless, R.I. = 1·868) and it has crystallized in the matrix environment and is not fragmental.

In one specimen the matrix material comprises about 60 per cent of the rock and consists of an equigranular aggregate (0·015 mm. grain size) of quartz with some antigorite laths or fragments. The matrix has comb structure developed along channels and marginally to some fragments.

(*d*) *Peridotite Breccia.*—The fragments in this type of breccia are characteristically unserpentinized and olivine (Fo_{95}; $\beta = 1{\cdot}662$) and enstatite remained stable phases under the conditions of brecciation. Olivine, enstatite and uncommon chromite fragments occur in sizes to less than 0·05 mm. and are typically angular. Larger peridotite fragments show the normal variable stress effects but in the peridotite breccias, particularly the sheet occurrences, larger fragments are commonly less abundant and there is a high (60 per cent) proportion of finer (< 2 mm.) fragmental material.

In some specimens mesh-texture and bastite serpentine, antigorite and talc are completely absent; in others a few larger peridotite fragments show partial alteration to antigorite or more rarely to mesh-texture serpentine. Talc is present in some breccias, usually replacing enstatite rather than olivine, and in some cases appears to form during the brecciation process, whereas in other cases it may be of earlier formation, accompanying the deformation effects in the peridotite.

The matrix material is similar to that present in the antigorite breccia and consists of chalcedonic and opaline silica, a fine fibrous serpentine mineral (rarely well crystallized and a very minor constituent in some breccias), and patches of anhedral magnesite, particularly common in

* Matrix serpentine mineral—pale yellow, non-pleochroic, fibrous to micaceous habit; positive elongation; biaxial, 2V low, optically negative; $\gamma = 1{\cdot}559$, $\alpha = 1{\cdot}547$; $\gamma - \alpha = {\cdot}012$, birefringence moderate.

the sheet breccias. Magnesite is absent in some breccias and in one breccia there is colloform garnierite in the matrix and in small veins penetrating peridotite fragments in the first stage of brecciation.

(*e*) *Gabbro Breccia.*—In the Wowo Gap area where the transgressive ultramafic breccias are most common the country rock is almost entirely of peridotite and dunite. However, further west there are gabbroic and transitional rocks in the Papuan Ultramafic Belt and in this area there are at least two small transgressive bodies of breccia within gabbro. The breccias consist of angular gabbro, bytownite and diallage fragments of very variable sizes in a turbid grey matrix probably mainly of fine saussuritized feldspar and some chalcedony. The breccias resemble the more common peridotite breccia bodies to the east in containing angular fragments of the immediate country rock with little or no admixed material and no evidence of shearing and tectonic brecciation.

(*f*) *Basalt-Peridotite Breccia.*—At Wowo Gap there is an area of breccia containing similar proportions of basalt and peridotite fragments. The breccia appears to be of the transgressive type although its boundary relations are uncertain, and probably marks a centre of the most intense brecciating activity.

The peridotite fragments are usually partly or completely altered to antigorite but magnesian olivine, enstatite, mesh-texture and bastite serpentine, and talc all occur in various fragments. Larger fragments show some degree of rounding. There are several fragments of quartz-microdiorite and related feldspar porphyry, also very rare fragments of quartzite, quartz-biotite and quartz-chlorite metamorphics. In similar proportions to the ultramafic fragments, and greatly in excess of the other very uncommon rock types, are fragments of basalt.

The basalt is an olivine alkali basalt with common phenocrysts of olivine, fewer and generally smaller phenocrysts of augite and a groundmass of zoned labradorite-bytownite laths with granular augite and magnetite and interstitial colourless glass. Modal nepheline is absent although nepheline is present in the norm (Table 1). Flow alignment of feldspar laths is commonly evident and in some fragments there is a distinct decrease in grain size near the fragment margin although there is no evidence of strong and rapid chilling. The olivine phenocrysts (Fo_{78}; $\beta = 1{\cdot}697$) characteristically show marginal alteration to bowlingite and in many fragments both glass and feldspar laths have altered to an extremely fine aggregate clay mineral. The character of the basalt is variable in its primary features in different fragments and in this differs from the peridotite fragments. This variation in the basalt is shown in the proportion of olivine phenocrysts, the proportion of augite, the transparency of the glass and the occurrence of magnetite as discrete crystals or as fine dust. There are also

textural variations, particularly in the presence or absence of flow structure. Some basalt fragments are almost spherical, many are finely vesicular and one example has a very irregular margin and several inclusions of breccia matrix material—these may be sections of breccia-filled pipe vesicles.

The matrix material is inhomogeneous; locally it is essentially a basalt crystal tuff with common olivine and bowlingite crystals, magnetite, pyroxene, feldspar (usually altered to a clay aggregate) and fine isotropic material (silica or glass). The olivine crystals are derived from the basalt and show no strain effects but commonly have a marginal bowlingite rim. In other areas the matrix has a more greenish colour and finely comminuted serpentine is an important constituent. Colloform chalcedony is locally present. Flow structure in the matrix is present in some places and is due to some parallelism of elongate fragments and a streakiness due to slight differences in the matrix material.

In summary the basalt-peridotite breccia differs from the previously described types in containing a mixture of fragments of unrelated rocks. This mixing is also evident in the variation in the type of alteration of the ultramafic fragments although antigorite is the dominant and thus probably the stable phase in the breccia. The basalt fragments appear to have been incorporated into the breccia before crystallization was complete whereas fragments of other rock types are of previously consolidated, and in some cases stressed and metamorphosed, rocks undergoing alteration due to inclusion within the breccia. The matrix material is essentially a basalt crystal tuff with admixed finely comminuted serpentine and some silica.

In two other breccias from the Wowo Gap area there are a very few fragments of hornblende andesine porphyry in which the andesine has largely been replaced by a clay mineral aggregate. These fragments are most probably derived from andesitic dykes which intruded the Papuan Ultramafic Belt and younger rocks following the deposition of the Domara River Beds. In one of these breccias the matrix contains common small bowlingite and magnetite grains as in the tuffaceous matrix of the basalt-peridotite type. In this breccia also the ultramafic fragments are largely of antigorite.

3. *The Occurrence and Inter-relationships of the Breccias.*—The ultramafic breccias have been divided into five types on their petrography and have two distinct modes of occurrence. The basalt-peridotite breccia is not known to occur as a sheet deposit. The sheet breccias are typically of the peridotite breccia and usually contain a large proportion of silt- and sand-size fragments with larger fragments of 1–3 inches average diameter. Peridotite breccia in Wowo Gap and particulatly south of Silimidi is of the transgressive type and shows a much greater size range in fragments than in the sheet breccias.

The antigorite breccia and the mesh-texture serpentine breccia are typical of the transgressive breccia bodies rather than the sheet occurrences. However, antigorite breccia occurring in Sivai Creek is probably derived from the base of a sheet breccia horizon. In some of the sheet breccias there are small channels and irregular patches in which fragments are of mesh-texture and bastite serpentine; these are interpreted as channels along which the final escape of fluids occurred during cooling of the sheet breccia. This could cause a lower temperature alteration of the fragments.

One specimen of ultramafic breccia consists of mesh-texture serpentine breccia as fragments within antigorite breccia. In this case there is a sharp contact between the two matrices which differ slightly in colour and transparency, and in coarseness of the matrix serpentine mineral. In the antigorite breccia there is some flow alignment of small fragments against the mesh-texture serpentine breccia fragments. In this example fragments of an earlier breccia formed by lower intensity brecciation have been caught up in a later phase of more intense activity forming the antigorite breccia.

4. *The Relationship of the Sheet Breccia to Sediments of the Silimidi Beds.*—The sheet breccias, consisting predominantly of the peridotite breccia type, occur as conformable horizons within the Silimidi Beds. No exposures of the base of the sheets were seen but near Silimidi is an exposure showing the upper contact of a sheet breccia with overlying sandstone. Lower in the sheet the rock is a normal peridotite breccia with a very few fragments of talc and antigorite. This type persists to about 6 inches below the upper surface of the sheet and at this level the chalcedony and serpentine mineral in the matrix have a colloform, encrusting habit around and separating fragments. In the upper 6 inches of the sheet there is a greater proportion of pale green matrix material; this contains patches and channels of magnesite but is largely green to yellow-brown, very fine chalcedony and serpentine mineral with aggregate and fibrous structure, locally becoming colloform around vughs. Many fragments have altered to nearly opaque, limonitic serpentinous material with some clear antigorite laths; other fragments contain antigorite with some talc and some fragments have altered to mesh-texture serpentine. The upper 6 inches of the sheet breccia thus corresponds more to the antigorite breccia than to the peridotite breccia type.

The top of the sheet breccia beneath the overlying coarse sandstone is a sharp bur irregular, non-planar boundary in hand specimen. There is no shearing parallel to the contact and the overlying sandstone is soft and friable, only slightly lithified and unmetamorphosed. In thin section the overlying sandstone has a greenish brown matrix of very fine fragmental material, in part fibrous but much less abundant than the

matrix of the sheet breccias and nowhere showing a colloform habit. The larger grains of the sediment average 1–3 mm. grain size and are commonly well-rounded. Serpentinite, peridotite and talc grains occur most commonly but there are also many crystal fragments of green hornblende and some of biotite. Carbonate (? magnesite) occurs as composite and single crystal fragments and rarely as small patches of cement. Most significantly there are a few grains with ultramafic breccia attached to serpentinized peridotite fragments and a few clastic grains of ultramafic breccia alone. The field and thin section evidence shows that the sheet peridotite breccias were deposited as conformable units in a sedimentary sequence and moreover were deposited as a single unit and not built up grain by grain as a sedimentary horizon. There may have been slight erosion of the upper surface of the sheet before deposition of the overlying sediment and deposition was probably terrestrial and fluviatile or possibly lacustrine rather than marine.

A thin section of a coarse sandstone from below the sheet breccias but within the Silimidi Beds shows that the most abundant fragments are of peridotite and derivatives but amphibolitized basalt and dolerite also occur. The green hornblende crystals in the altered basalt and dolerite are also present as single crystal fragments.

5. *Chemical Composition of the Breccias.*—In Table 2 ten chemical analyses of fragments and matrix in four types of breccia are given. The matrix and fragments were separated by hand crushing and although larger fragments could be freed of matrix material the analysed matrix material still contained the smallest of the fragments (<1 mm.) and so does not correspond only to material of direct matrix crystallization.

The analyses of the least serpentinized fragments (F_1 and F_3) from the peridotite breccias show that chemically these are a typical dunite (F_1) and enstatite olivinite (F_2) of the alpine ultramafic suite (Hess, 1938). The basalt within the basalt-peridotite breccia is an olivine basalt of the alkali magma suite. It is to be noted that in the basalt and the dunite both the modal olivine compositions (Fo_{78} and Fo_{95} respectively) and the normative olivine compositions (Fo_{77} and Fo_{95} respectively) are very different in the two rock types.

The pairs of analyses of fragments and matrix show that in the process of brecciation the compositional changes to give the matrix composition from the original fragment composition are: (1) Addition of the volatiles H_2O, CO_2 and possibly Cl and S.

(2) Addition of SiO_2 (or removal of MgO) with decrease in the MgO/SiO_2 ratio. SiO_2 and CO_2 have to some extent an antipathetic relationship in the matrix but from thin section examination it seems that the distribution of CO_2 is very patchy and haphazard and

TABLE

	M_1	F_1	M_2
SiO_2	42·09	40·44	33·49
Al_2O_3	1·90	0·76	2·25
Fe_2O_3	1·59	1·82	1·85
FeO	3·78	5·01	2·23
MgO	43·60	48·39	43·07
CaO	* trace	* trace	0·08
Na_2O	0·02	0·02	0·02
K_2O	0·02	0·03	0·02
H_2O^+	2·74	1·94	6·91
H_2O^-	2·46	0·51	2·28
CO_2	0·16	0·08	6·82
P_2O_5	0·05	0·06	0·05
TiO_2	Nil	Nil	Nil
MnO	0·11	0·11	0·02
Cr_2O_3	0·35	0·41	0·31
NiO	0·81	0·82	0·30
Cl	0·05	n.d.	0·05
S	0·03	n.d.	0·04
	99·76	100·40	99·79
Total Fe as Fe_2O_3	5·78	7·39	4·33
Oxidation Ratio $\frac{Fe_2O_3}{\text{Total Fe as } Fe_2O_3}$	0·275	0·246	0·42
$\frac{MgO}{SiO_2}$ mol. ratio (less MgO as $MgCO_3$)	1·54	1·78	1·64
$\frac{Mg}{Fe}$ mol. ratio		13·0	

* CaO < 0·05%

M_1 Matrix of specimen 85941 } Slig
F_1 Fragments of specimen 85941 }
M_2 Matrix of specimen 85942 } Slig
F_2 Fragments of specimen 85942 }
M_3 Matrix and small fragments [< 1 cm
F_3 Basalt fragment from specimen 8594
M_4 Matrix of specimen 85944 } Ant
F_4 Fragments of specimen 85944 }
M_5 Matrix of specimen 85945 } Mes
F_5 Fragments of specimen 85945 }

Norm F_1—dunite [very slight serpentinization]

Olivine	Forsterite	79·
	Fayalite	6·
Orthopyroxene	Enstatite	8·
	Ferrosilite	0·
Spinel	Spinel [$MgAl_2O_4$]	1·
	Chromite	0·
	Magnetite	2·

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TABLE 2.—CHEMICAL ANALYSES

M_2	F_2	M_3	F_3	M_4	F_4	M_5	F_5
33·49	38·57	42·91	44·42	40·41	40·20	41·11	39·10
2·25	1·02	10·17	15·49	1·39	2·04	1·01	1·08
1·85	2·61	4·50	3·05	3·99	3·59	4·98	5·35
2·23	4·11	2·37	6·92	1·10	3·70	0·92	2·27
43·07	46·70	23·11	12·13	37·68	42·83	36·39	42·06
0·08	0·24	2·46	9·78	* trace	* trace	0·42	0·28
0·02	0·02	0·78	2·31	0·01	0·00	0·01	0·00
0·02	0·02	0·65	1·49	0·02	0·01	0·01	0·00
6·91	5·08	7·74	1·70	10·01	6·63	11·50	8·19
2·28	0·06	4·68	0·84	3·59	0·48	2·90	1·28
6·82	0·88	0·09	0·06	1·13	0·06	0·29	0·16
0·05	0·04	0·04	0·05	0·04	0·03	0·04	0·03
Nil	Nil	0·64	1·15	Nil	Nil	Nil	Nil
0·02	0·10	0·15	0·15	0·03	0·05	0·05	0·05
0·31	0·34	0·20	0·07	0·22	0·27	0·35	0·37
0·30	0·49	0·30	trace	0·75	0·61	0·71	0·45
0·05	n.d.	0·04	0·10	0·10	n.d.	0·03	n.d.
0·04	n.d.	0·03	0·11	0·06	n.d.	0·02	n.d.
99·79	100·28	100·86	99·82	100·53	100·50	100·74	100·67
4·33	7·18	7·14	10·74	5·21	7·70	6·01	7·88
0·428	0·364	0·630	0·284	0·765	0·466	0·828	0·679
1·64	1·78	0·81	0·41	1·35	1·59	1·32	1·60
	12·9		2·24		11·0		10·6

Analyst : D. H. GREEN

} Slightly serpentinized transgressive peridotite breccia with garnierite

} Slightly serpentinized sheet periodotite breccia

< 1 cm.] of specimen 85943 } Basalt-peridotite breccia
en 85943

} Antigorite serpentinite breccia

} Mesh-texture serpentinite breccia

ization]

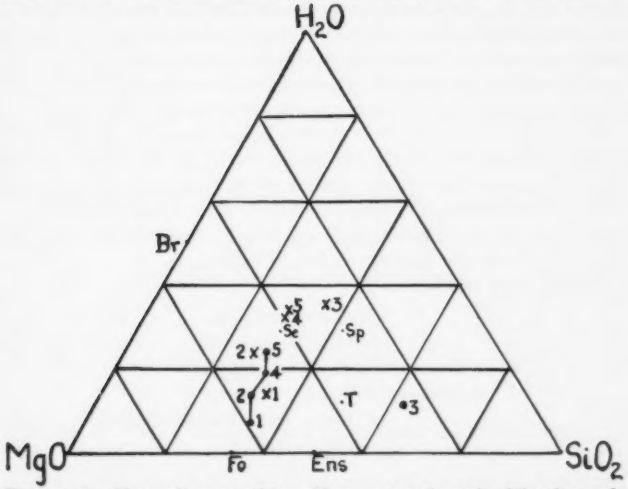
79·8		
6·7	} 86·5	Olivine Fo_{85}
8·5		
0·5	} 9·0	Enstatite En_{93}
1·1		
0·6	} 4·3	Spinel
2·6		

Norm F_5—olivine alkali basalt

Orthoclase		9·4	
Nepheline		4·1	
Albite		10·4	} 40·2
Anorthite		29·8	
Diopside	Ca	8·6	} 16·4
	Mg	6·0	
	Fe	1·8	
Olivine	Forsterite	17·7	} 23·6 (Fo_{77})
	Fayalite	5·9	
Magnetite		4·6	
Ilmenite		1·4	
		99·7	

high CO_2 cannot be regarded as characteristic of one type of breccia or mode of occurrence.

(3) Addition of Al_2O_3. The matrix in the peridotite breccias is significantly richer in Al_2O_3 than the fragments. In the basalt-peridotite breccia there is also an enrichment in Al_2O_3 in the matrix to a greater extent than that predicted from simple mechanical mixing of basalt and peridotite fragments in the proportions set by the other oxides such



TEXT-FIG. 3.—Change in composition of fragments and matrix of the ultramafic breccias : • = fragments, × = matrix, numbers as in Table 2. Br, brucite ; Se, serpentine ; Sp, sepiolite ; T, talc ; Fo, forsterite ; En, enstatite.

as TiO_2, CaO, Na_2O, K_2O. However, in the antigorite breccia the converse relationship holds and it appears that in the antigoritization process Al_2O_3 enters into the mineral phases of the fragments rather than remaining in the matrix. In the mesh-texture serpentine breccia the Al_2O_3 content is approximately the same in both matrix and fragments.

(4) Relative removal of MgO, total Fe, and Cr_2O_3. These constituents are probably diluted rather than removed by the " volatiles " H_2O, CO_2, SiO_2 and Al_2O_3.

The changes in bulk composition from fragments to matrix (Table 2 and Text-fig. 3) are paralleled by changes in the state of oxidation (Text-fig. 4) so that oxygen must also be regarded as being added.

It is stressed that the matrix material is not of fixed composition in the different breccias but is of an arbitrary composition somewhere

between that of the fragments (whether of one or more rock types) and a variable mixture of H_2O, SiO_2, CO_2, O_2 and Al_2O_3. There is no suggestion that a magma of definite composition (i.e. that of serpentine cf. Hess, 1938 and Wilkinson, 1953) forms the true matrix in each case.

The mineral alteration of the fragments discussed previously seems to be directly attributable to the brecciation process and the chemical changes in the fragments generally parallel those in the matrix (Text-figs. 3 and 4). Thus there is an increase in H_2O and O_2 content, a removal of MgO and a relative increase in SiO_2 and in one case an addition of Al_2O_3. However, there is no significant addition of CO_2 and no relative removal of Cr_2O_3 and total Fe and in these respects the chemical changes in the fragments differ from those in the matrix.

6. *Stability Relationships of the Serpentine Minerals.*—Three of the types of ultramafic breccia are distinguished on the nature of the particular magnesium silicate present in the fragments. The breccias show arrested stages of conversion of these three types one to another and from the thin sections it is possible to recognize reaction and equilibrium relations between certain mineral pairs.

Replacement relations:—

olivine → antigorite
enstatite → antigorite
mesh-texture serpentine → antigorite
bastite serpentine → antigorite
olivine → mesh-texture serpentine
enstatite → bastite

There is no evidence in the breccias of reversibility in these reactions.

Equilibrium assemblages:—

(*a*) olivine + enstatite + spinel.
(*b*) antigorite (+ chromite ? + magnetite ?).
(*c*) mesh-texture and bastite serpentine + magnetite or haematite.

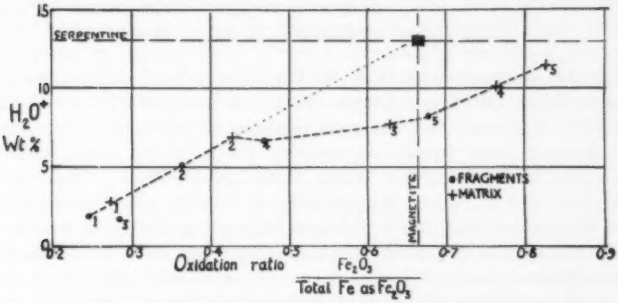
Talc is not a major constituent in any of the breccias examined so its position in these assemblages is not certain. However, in some cases it seems to be a replacement of enstatite and in equilibrium with olivine, in others a replacement of enstatite and in equilibrium with antigorite.

From the examination of the breccias the following conclusions can be drawn:—

(1) Under low intensity brecciation of the peridotite, mesh-texture and bastite serpentine, and fibrous chrysotile were stable and replaced the primary olivine and enstatite. Antigorite, if formed at higher intensities of brecciation, remained stable under the lower intensity conditions but olivine and enstatite altered readily. There is some variation in colour and birefringence of the mesh-texture serpentine

due to retention of iron in the structure or its exsolution as very fine dust or as larger magnetite or haematite grains.

(2) Under higher intensity conditions antigorite was the stable phase; olivine, enstatite, mesh-texture and bastite serpentine were all unstable and completely or partly replaced by antigorite. Antigorite commonly formed from mesh-texture serpentine with resorption of exsolved magnetite. In the lower part of the stability field mesh-texture serpentine survived as opaque or very turbid relics in which the mesh or window structure has largely been destroyed by growth of some portions of the mesh, usually the opposite sides, at the expense of the others. This form of serpentine appears to be a metastable form



TEXT-FIG. 4.—Plot of oxidation ratio against combined water content of the ultramafic breccias : • = fragments, + = matrix, numbers as in Table 2.

developed prior to replacement by clear antigorite laths in random arrangement.

(3) Under the highest intensity conditions [the formation of the sheet and vent peridotite breccias] olivine and enstatite remained stable in an environment with free SiO_2, H_2O and CO_2.

(4) The conditions of final crystallization of the matrix material were such that hydrous, low temperature minerals crystallized, commonly in colloform habit. Chrysotile, chalcedony and magnesite are normal direct crystallizations in the matrix and more rarely garnierite, andradite and quartz have crystallized directly in the matrix in different examples.

The stability relationships observed in the breccias should be interpretable with reference to Bowen and Tuttles' (1949) experimental work on the system MgO-SiO_2-H_2O. The peridotite breccias in which olivine and enstatite are unaltered even in the finest fragments would probably have been formed at a temperature above 600° C. (where no talc is present) and certainly above 500° C. (but less than 600° C. where

talc has partly replaced enstatite). The mesh-texture serpentine breccias in which olivine and enstatite have altered directly to mesh-texture serpentine and bastite have approached a stable mineral assemblage at less than 500° C. and most probably at less than 400° C. (reaction forsterite + water $\rightleftarrows$ brucite + serpentine).

In Bowen and Tuttles' work chrysotile was the only serpentine mineral formed and thus their results cannot be strictly applied to natural rocks in which three forms of chrysotile, lizardite and antigorite may all occur (Whittaker and Zussman, 1956). Of particular importance in the petrogenesis of the ultramafic rocks is the fact that there are no experimental data on the reactions involving antigorite e.g.

olivine + enstatite $\rightarrow$ antigorite
mesh-texture serpentine + bastite $\rightarrow$ antigorite

From observational data on the the genesis of antigorite in serpentines (Hess, Smith and Dengo (1952); Francis (1956); Wilkinson (1953)) it appears that antigorite replaces mesh-texture and bastite serpentine under conditions yielding rocks of the albite-epidote amphibolite metamorphic facies. Some authors (Hess, Smith and Dengo, 1952, p. 74) have suggested that it forms only under conditions of dynamothermal metamorphism and that directed stress is necessary or favourable to its growth. The growth of antigorite in the ultramafic breccias here described is a clear example in which directed stress has played no part and the physical variables have been temperature and partial and total gas pressures.

The chemical analysis of the partially antigoritized fragments in the antigorite breccia shows that these contain twice as much Al_2O_3 as is present in the unserpentinized peridotites or in the mesh-texture serpentine. Assuming all the H_2O^+ and all the Al_2O_3 of the analysis are in antigorite then the antigorite would contain approximately 3·7 per cent Al_2O_3. This is lower than but not incompatible with the average Al_2O_3 content (6·5 $\pm$ 2·9 per cent) of antigorites given by Nagy and Faust (1956). An observed feature of antigoritization of the fragments is that where antigorite replaces mesh-texture serpentine it usually does so with reabsorption of exsolved magnetite. Similarly the antigoritization of olivine and enstatite may occur without exsolution of magnetite or with lesser exsolution than occurs in the similar replacement by mesh-texture serpentine. These observations imply that antigorite contains Al^{+++} and Fe^{+++} with the Fe^{++} substitutions in the usual serpentine formula ($Mg_3 Si_2 O_7 \cdot 2H_2O$) and probably these are necessary for stability of the structure. This substitution probably accounts for some of the features of fig. 4. For most of the chemical analyses there is a linear interdependence between the degree of hydration and the degree of oxidation. This is explained by the fact

that as the olivine and enstatite are serpentinized the ferrous iron within them is exsolved as magnetite or haematite. However, for two analyses, M_3 and F_4, the degree of oxidation is independent of the degree of hydration. In both of these antigorite is abundant. If antigorite contains Fe^{+++} substitutions then the degree of oxidation of the iron within the rock may vary without change in the degree of hydration.

The role of antigorite in the ultramafic rocks needs clarification, particularly with respect to its range of composition and to its stability field. From previous petrographic studies and from the present it is clear that antigorite forms at higher temperatures than mesh-texture serpentine and bastite (which are in this case chrysotile-lizardite mixtures) but is stable under both regional metamorphic and low-pressure, thermal conditions as in this example. In addition it is most probable that the presence of Al^{+++} and/or Fe^{+++} is necessary in the rock for the formation of antigorite and antigorite is not a stable phase in the simpler system MgO-SiO_2-H_2O.

Genesis of the Ultramafic Breccias

In deducing the origin of the ultramafic breccias the processes which must be considered may be grouped as (1) Tectonic (2) Volcanic or magmatic (3) Sedimentary (4) Talus brecciation (5) Solution brecciation. Any process invoked must be capable of producing both modes of occurrence of the breccias and of producing the features of their petrography and chemistry detailed previously. The only process which seems capable of doing this is some process of volcanic brecciation. Tectonic brecciation is inconsistent with the sheet breccias and particularly with the lack of shearing effects within the matrix of the breccias. Talus and solution brecciation are also inconsistent with the features of the sheet breccias, the sequence of mineral alteration and the textures of the breccias. Sedimentary processes could not explain the transgressive breccias or the lack of admixed material and the textural features in the sheet breccia.

Volcanic processes could explain the transgressive and sheet occurrences of the breccia as vent and effusive deposits of the one process. Two possibilities exist: (1) The rocks are vent breccias and effusive volcanic breccias produced by an ultramafic magma, or, (2) the rocks are vent breccias and extruded vent breccias of peridotitic country rock brecciated and, in the sheet occurrences, entrained by volcanic gases derived from a subjacent source of no genetic relationship to the ultramafic-basic suite.

The first possibility may be excluded. The ultramafic fragments are fragments of previously consolidated, emplaced and dynamically metamorphosed rocks that show secondary alterations due to inclusion

within the breccia. Also there is a large time break between emplacement of the ultramafic-basic suite of rocks and the activity producing the ultramafic breccias so that genetic relation of these two seems less likely. The matrix material does not have a constant composition suggesting an ultramafic or serpentine magma and petrographic and chemical evidence shows that the constituents added to the previously emplaced peridotite during the brecciation process are H_2O, SiO_2, Al_2O_3, CO_2 and O_2. The minerals of direct crystallization in the matrix (chrysotile, chalcedony, quartz, magnesite, garnierite, andradite) and their textures are inconsistent with normal magmatic crystallization, even of an ultramafic magma, but are consistent with a hydrothermal environment.

The second hypothesis is considered to be the only one consistent with the whole of the observations presented. The process envisaged is that volcanic gases from a subjacent source (magma) penetrated and brecciated the country rock of dunite and peridotite. The sequence of breccia types, mesh-texture serpentine breccia, antigorite breccia and peridotite breccia, is consistently explained as due to increasing intensity of brecciation correlated with increasing temperature and gas pressure of the penetrating gases. The nature of the gases was probably near that of normal volcanic gases (Shepherd 1938; Krauskopf 1959) and predominantly gas-phase water with CO_2, O_2, and probably with some SiO_2 and Al_2O_3, in solution. The minerals and the texture of direct matrix crystallization appear to be consistent with deposition from this environment during cooling of the penetrating fluids.

The transgressive breccias are interpreted as being the pipes and channels within the country rock along which escape of the gases occurred and the variation from finely jointed peridotite through matrix-poor serpentine breccia to matrix-rich peridotite breccia seen in some bodies is related to increasing gas pressure and velocity. The gabbro breccia is the result of the same process acting on gabbroic country rock rather than peridotite.

The process of formation of the vent breccias i.e. gaseous penetration of a fragmented rock, is analogous to the industrial process of fluidization. The sheet breccias are considered to be the overflow and extrusion from one or more vents of moderate to high temperature fluidized ultramafic breccia. As such they would correspond with the entrained stage of the industrial fluidized system where the gas pressure and velocity becomes high enough to transport the liquid-like, fluidized system as a whole. As would be expected the extruded breccia was of the peridotite type i.e. the type formed in the vents of highest intensity of brecciation.

The sheet breccias are very similar in origin to nuée ardente deposits in that high temperature, high pressure volcanic gas produces the

fluidity of a suspension of rock fragments and dust. However, they differ from nuée ardente deposits in that none of the material was liquid at the time of extrusion and none of it is the parental magma providing the volcanic gases. The nuée ardentes of Mt. Lamington, Papua (Taylor, 1958) transported much fragmental country rock (older lavas) but also contain a proportion of new lava fragments and single crystals. In lacking any of the volcanic material parental to the gases the ultramafic sheet breccia resembles the intrusive " lithic tuff " described by Francis (1959) and several of the vent breccias described by Reynolds (1954, p. 584, p. 587). The breccias differ from " tuffisites " (Reynolds, 1954; Whitten 1959) in not containing acid, aplitic material as the fluidized rock. The writer is not aware of any described examples in which such " lithic tuffs " and vent breccias of country rock alone have formed nuée ardente-type extrusions giving agglomerate-like deposits, particularly of the extent of the Musa Valley occurrences.

In the basalt-peridotite breccia attention has been drawn to the fact that much of the matrix material approaches a basalt-tuff. Also, although the ultramafic fragments have the characters of a previously consolidated rock type caught up in the brecciation process, the basalt fragments were incompletely crystallized when fragmentation occurred. It thus seems likely that the source of the volcanic gases lay in a subjacent olivine alkali basalt magma, and that locally, at the apparent centre of most intense activity at Wowo Gap, some of this magma became intermixed with the country rock in one of the vents. Corroborative evidence that such volcanic activity occurred in the province is found in the penecontemporaneous Sesara Volcanics (olivine basalt) outcropping about 30 miles N.W. of Wowo Gap. It is possibly relevant that the two recent eruptions in the province (Waiowa or Goropu in 1943–44 and Mt. Lamington in 1951) were predominantly effusive and gave rise in both cases to nuée ardentes (Taylor, 1958). These volcanics differ from the alkali basalts of Wowo Gap and chemically may be classed as doreites (Mt. Lamington) and possibly latite (Waiowa) on the basis of the average chemical compositions given by Nockolds (1954). Both rock types contain phenocrysts of augite, biotite and brown hornblende.

The Significance of the Papuan Ultramafics to the Serpentinite Problem

The major controversy in the genesis of the alpine ultramafic rocks revolves around the existence or non-existence of a low temperature ultramafic magma at shallow to moderate crustal depths. The two main conflicting views have been well presented by Bowen (1949) and Hess (1938, 1955) and the problem has been reviewed and restated with

relevant evidence by Ross *et al.* (1954), Wilkinson (1953) and Ruckmick and Noble (1959) among others. The features of the Musa Valley outcrops of the Papuan Ultramafic Belt described in this paper are very relevant to the whole problem of the alpine ultramafics.

In modifying Benson's (1926) classification, mainly environmental, of the ultramafic rocks Hess (1938, 1955) defined the alpine ultramafic suite in terms of its characteristic environment, chemistry, petrography and associated rock types. Ross *et al.* (1954) have also given petrographic and chemical data on dunites of the alpine suite. The peridotites and dunites of the Musa Valley area are typical members of the alpine ultramafic suite (Hess, 1938) in their environment, their chemistry (particularly their high Mg/Fe molecular ratios) and in their petrography (particularly in the modal olivine composition and the strong stress effects). However, they equally clearly form an integral part of a differentiated, layered rock series in which bytownite gabbro is a major rock type and which contains interlayered transitional rock types.

The Musa Valley suite appears to have affinities with inclined and folded, layered ultramafic and mafic rocks occurring in alpine environments in the Zambales Range, Philippines (Gonzales *et al.*, 1957); Camaguey district, Cuba (Flint *et al.*, 1948); Troodos Range, Cyprus (Gass, 1958; Wilson, 1959) and possibly the Bay of Islands, Newfoundland (Ingerson, 1935; Smith 1953). In other alpine ultramafics compositional banding attributed to gravitational differentiation may be more fragmentary (Wilkinson, 1953 p. 311; Little, 1949, p. 808). In the Philippines the basic and ultramafic rocks occur in alternating belts 4–9 miles wide; both rock groups are rhythmically layered and some contacts are interlayered.

Although the data are meagre, particularly in the petrographical and chemical aspects, ultramafic and basic accumulates of the Musa Valley and Philippines type (and probably more generally of the "alpine type") appear to differ from accumulates from a basaltic magma i.e. Bushveld, Skaergaard, Stillwater and Rhum Complexes, in the following features:—

1. The very great thickness of the basic, transitional, and ultramafic groups.

2. The apparent repetition of such units without a clearly defined sequence from accumulative ultramafic to residual granophyric rocks.

3. The absence of rocks more acid than bytownite gabbro.

4. The high Mg/Fe molecular ratio of the ultramafic accumulates (Hess, 1938).

5. The highly magnesian olivine (Fo_{95}) and enstatite in the dunites and peridotites contrasting with the olivines from early accumulates

of basaltic magmas (Stillwater Fo_{87}; Bushveld Critical Zone Fo_{88}; Rhum Fo_{86}).*

An additional complexity is the difficulty of envisaging a magma chamber of appropriate dimensions undergoing prolonged differentiation in the active orogenic environment in which the ultramafic rocks occur.

In the Musa Valley no intrusive contacts of the ultramafic-basic suite with sedimentary or metamorphic rocks were seen, The folded nature of the layered series, the strong stress effects in the peridotites and dunites and the faulted and sheared contacts suggest that the ultramafic-basic series may have been tectonically emplaced at its present crustal level, and that this environment is irrelevant in the discussion of the primary differentiation and accumulation of the layered series. Ross *et al.* (1954) have argued in favour of a sub-crustal derivation of alpine dunites and peridotites and of olivine and related nodules in basaltic rocks. Applying their conclusion to the Papuan ultramafics and to the other similar examples referred to above one is led to the hypothesis that either the sub-crustal levels may be themselves layered in the manner of gravitationally accumulative igneous rocks or alternatively that such levels frequently contain large magma chambers of basaltic to ultramafic composition in which very thick accumulative sequences may form. The ultramafic rocks of the Musa Valley provide no positive support for the concept of the existence of a peridotitic or dunite magma at *crustal* levels. The distinction between ultramafic rocks of Bushveld and alpine type (Hess, 1938) probably remains a valid genetic distinction but the criteria for classification given by Hess must be modified in view of the layered nature of the Papuan and other ultramafics and of their primary origin as accumulates from a magma of uncertain composition.

In advocating the existence of a low-temperature, water-rich primary peridotite magma approximating to the composition of serpentine, Hess (1938) suggested that volcanic activity due to such a magma would be characterized by effusive activity with escape of large volumes of gas. Hess considers that ultramafic breccias in northern Italy and in south-eastern Borneo may represent this type of activity.

Wilkinson (1953) has suggested that serpentinized ultramafic breccias from Queensland are due to the intrusion and brecciation of earlier peridotite bodies by a later magma approximating to serpentine in composition. The breccias described by Wilkinson differ from the Papuan examples particularly in their complete serpentinization and in the schistose nature of much of the fine matrix material. However, both these differences may simply reflect the effect of the more complex

* The olivines of the Great Dyke, Southern Rhodesia (Worst, 1958) range from Fo_{94} to Fo_{84} and may provide a link between the suites.

history and greater age (Lower Carboniferous) of the Queensland breccias. Any later general serpentinization would tend to blur the primary features of the Papuan breccias and later stress effects would be first evident in the matrix of these breccias so that a change to breccias practically identical with the Queensland examples can be readily envisaged. The analysis of the matrix material compared with that of the associated fragments (analyses 1 and 2, Wilkinson, 1953) shows similar changes to those in the Papuan breccias, i.e. higher H_2O, higher $\frac{SiO_2}{MgO}$ ratio and lower total Fe in the matrix material.

On first impressions the Papuan breccias seemed to offer good examples of the vent and effusive phase of the volcanic activity of an ultramafic magma. A more detailed investigation has shown this not to be so but rather that the peridotite has played the passive role of country rock to an unrelated volcanism. Since large areas of ultramafic rocks are common in young orogenic environments and repeated volcanism is equally characteristic of such environments then the process of formation of the Papuan ultramafic breccias could well be repeated elsewhere. In particular the occurrences of ultramafic breccia mentioned above could possibly be interpreted by this hypothesis as an alternative to that relating them to a hydrous, primary peridotite magma.

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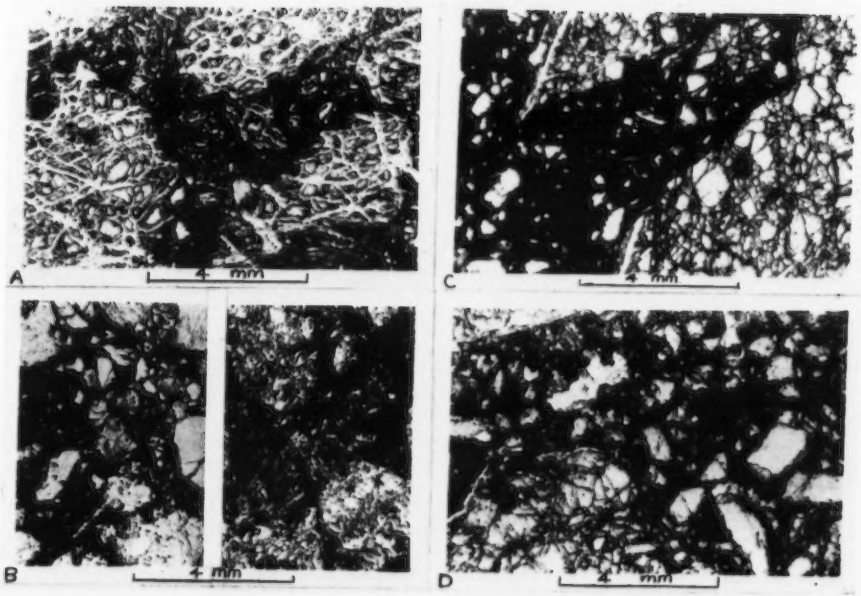
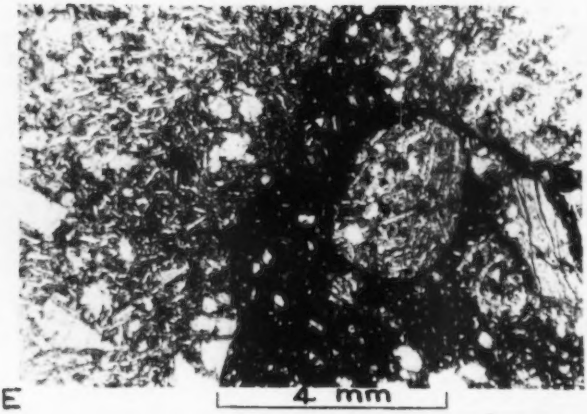


ULTRAMAFIC BRECCIAS FROM PAPUA.

Facing page 24



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EXPLANATION OF PLATES

PLATE I

A.—Mesh-texture serpentine breccia ; dunite partly replaced by mesh-texture serpentine, finer fragments of serpentine and olivine, turbid matrix of serpentine and silica with colloform structure in lower right corner. (Ordinary light.)

B.—Antigorite serpentine breccia ; fragments of random antigorite laths, in some cases with fine magnetite. (Left—ordinary light ; right—crossed nicols ; adjoining fields.)

C.—Peridotite breccia (transgressive type); brecciated dunite, fine-grained due to earlier crushing. Turbid matrix includes garnierite. (Ordinary light.)

D.—Peridotite breccia (sheet type); fragments of unserpentinized olivine and enstatite; patch of magnesite cement to left of centre and a cluster of chromite fragments to right of centre. (Ordinary light.)

PLATE II

E.—Basalt-peridotite breccia—large basalt fragment with fine-grained margin at left, antigorite serpentine at upper right and partly serpentinized peridotite (mesh-texture serpentine) at middle right. (Ordinary light.)

F.—Antigorite serpentine breccia. At right is a partly antigoritized peridotite fragment and diagonally across the centre is an area of well crystallized matrix minerals including clinochrysotile (clear and grey, distinctly fibrous) and andradite (high relief, deep grey, sub-hedral). (Ordinary light.)

Cross-bedding in the Calciferous Sandstone Series of Fife and West Lothian

By J. T. Greensmith

Abstract

Cross-bedding within the sandstones of eastern Fife indicates a predominant north to north-easterly source for the clastic debris. This inference is in accordance with cross-bedding data from sandstones of similar age in southern Fife and West Lothian.

I. Introduction

Despite their bearing on palaeogeographical problems such as the position of source areas and conditions of deposition (Potter and Siever, 1956), cross-bedding studies are rare in the British literature. Sorby (1859), Gilligan (1920), Stephens (1952) and Walker (1955), all examined Millstone Grit sandstones in Yorkshire and Derbyshire, and Sorby (1853) referred briefly to cross-bedding in the Edinburgh area. Schwarzacher (1953) studied Cretaceous sandstones in central England, and more recent work includes that of Robson (1956) on the Fell Sandstones and Knill (1959) on Dalradian meta-sediments.

The main area selected for this present study is East Fife and, for comparison, reference will be made to work published on southern Fife and West Lothian (Greensmith, 1960, in press). In East Fife there is exposed along the coast a nearly continuous succession of gently folded and faulted Calciferous Sandstone Series rocks (Kirkby, 1880, 1901, Geikie, 1902) in which cross-bedded sandstones are common. The sandstones, which are interbedded with shales, coal seams and thin limestones, have been examined from Buddo Ness, three miles east of St. Andrews, eastwards to Fife Ness, then southwards to Elie, a distance of about twenty miles (Text-fig. 1).

II. Cross-bedding in East Fife

The sandstones, which are predominantly ortho-quartzites with subordinate sub-greywackes, are relatively shallow-water deposits as lenticularity, ripple-marks and wash-outs are common features. Ripplemarks are well seen at Roome Harbour, Crail (Nat. grid ref. 621078) as are wash-outs, and there are further excellent examples of the latter on the foreshore at Kilrennymill (582044) and Anstruther (573035). It is, however, the lenticular bedded sandstones which most often exhibit cross-bedded structures, although it is not always possible to determine their three-dimensional attitude. This difficulty is due to " honey-comb " weathering as at Pan Haven, south of Crail (607067) or patchy post-depositional haematite staining which can simulate

EAST FIFE
BUDDO NESS
CAMBO
FIFE NESS
CRAIL
CAIPLIE
ANSTRUTHER
ST. MONANCE
ELIE
134 READINGS
GRAND MEAN
CARB. 151 SERIES
MILES
105 READINGS
GRAND MEAN
SOUTHERN FIFE AND WEST LOTHIAN

cross-bedding. Then, at innumerable localities such as Buddo Ness (563149), Almond Rock (611070), and Kilminning (632086) complicated slump-bedding effectively prevents a restoration of the fore-set beds to their original position. These slumped beds have been excluded from the study.

In the undisturbed beds elsewhere the corrected angle of dip of the fore-set beds is never greater than 15 degrees though, prior to compaction, it may have been slightly higher (Sorby, 1908). The fore-set beds are generally slightly concave downwards, sometimes linear and, very rarely, disposed in festoon fashion. Within any particular sandstone all three types may be present.

More important than the actual dip of the fore-set beds are their directions of dip simply because these indicate, in aggregate, the overall direction of clastic sediment transportation and, probably, the broad regional tilt of the basin of deposition. At each sandstone loçality, therefore, as many directional observations as possible have been made. The maximum number recorded at any one locality was five, but in many instances, only one or two could be made accurately because of the factors mentioned earlier.

Where two or more readings were possible the average dip directions have been calculated and it is these which are plotted in Text-fig. 1. For localities where four or more readings were possible the arithmetic standard deviation has been calculated also and is found to vary between 20 and 65 degrees.

Text-fig. 1 clearly indicates that the areal trend of fore-set bed average directions is by no means uniform and, in fact, a range through nearly 360 degrees is present. This is to be expected under the shallow water environmental conditions which presumably prevailed at the time of clastic deposition and, no doubt, reflects localized changes in current flow due, in part, to the irregularities of the basinal floor.

The wide directional range in bearings is made even more explicit when all the specific directional readings from all the loçalities are plotted in the form of a rose diagram. Altogether, 134 readings have been plotted in this fashion (Text-fig. 1) and grouped into 40 degree class intervals (azimuths). The area of each class is proportional to the number of readings within it. Apart from graphiçally indicating the 360 degree variation it is readily apparent that most of the readings fall into the 160 to 240 degree azimuthal range. Thus, it seems likely that by far the largest bulk of clastic material has been transported from a direction varying between north and north-east and this is an inference which is substantiated to some extent by the value of the grand arithmetic mean (M_a) based on all the readings. This mean value gives the overall direction of transport just towards the west of south at 190 degrees with an arithmetic standard deviation of $\pm$ 78 degrees

(Krumbein and Pettijohn, 1938). The azimuthal and statistical data are presented in Table 1.

TABLE 1

Eastern Fife

Azimuths	*No. of readings.*	*f*	*m*	$f(m - M_a)^2$
0—40	6	4·5	20	138,000
41—80	12	8·8	60	148,700
81—120	10	7·4	100	59,940
121—160	16	11·8	140	29,490
161—200	25	18·6	180	1,860
201—240	34	25·6	220	23,030
241—280	17	12·6	260	61,720
281—320	11	8·2	300	99,220
321—360	3	2·5	340	56,240
	134			610,200

$M_a = 190°$ $\sigma_a \pm 78°$

Southern Fife and West Lothian

0—40	1	0·9	20	31,460
41—80	4	3·8	60	82,130
81—120	8	7·6	100	87,020
121—160	17	16·1	140	72,280
161—200	23	21·9	180	15,960
201—240	16	15.2	220	2,568
241—280	17	16·2	260	45,360
281—320	13	12·4	300	107,300
321—360	6	5·9	340	104,300
	105			548,378

$M_a = 207°$ $\sigma_a \pm 74°$

III. Comparison with South Fife and West Lothian

Sediments of similar age, style of folding, and lithology, but less marine in character, are exposed in the southern part of Fife and West Lothian. There, as in Eastern Fife, cross-bedded sandstones are common and many of the fore-set beds show considerable internal contortion resulting from post-depositional movement. No measurements from these contorted beds have been considered. From the remainder 105 readings were made and the resultant information is plotted in similar form in Text-fig. 1. Once again the azimuthal distribution suggests a dominant clastic source area towards the north-east and partial confirmation comes from the grand mean direction of 207 degrees that is, approximately towards the south-south-west. The standard deviation (± 74 degrees) and relevant data for comparative purposes are recorded in Table 1.

IV. Conclusions

The new information suggests that these two now disconnected areas of clastic sedimentation were one during Calciferous Sandstone times. This, of course, is confirmed by normal stratigraphical considerations. Furthermore it is a unity which can be substantiated by other statistical methods such as the " coefficient of correlation " (Dryden, 1935). It is unusual to apply this particular method to cross-bedding studies, but a useful purpose is served in this case in mathematically relating the two sets of information to each other. Assuming the unlikely occurrence of identical fore-set bed reading distribution in both areas the coefficient would be + 1. Conversely, if the distribution was grossly dissimilar a very low coefficient would be obtained. As, in fact, the coefficient derived is 0·70 it is clear that a strong degree of similarity exists between the two sets of data and this accords with the more graphically expressed information in Text-fig. 1.

It is also clear that during Calciferous Sandstone times the input of clastic material at the eastern end of the Midland Valley area of sedimentation was dominantly from the north and north-east. The regional depositional slope possibly had a general tilt down to the south and south-west.

The actual source of the clastic material in eastern Fife remains conjectural until detailed petrographic work has been completed but a Lower Old Red Sandstone ridge has already been suggested for the southern Fife and West Lothian sandstones (Greensmith, 1960).

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On the Significance of β Intersections in Superposed Deformation Fabrics

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Abstract

The uncertainty of the most frequent orientations of β and fold axes relative to tectonic co-ordinates has possibly been over-emphasized in tectonic literature. The β concentrations of obliquely deformed folds will be spread along characteristic paths. The superposed deformation would in most cases of flexural slip appear as folds in certain predominant directions and on suitably oriented surfaces. In tectonites with a single, nearly horizontal mode for the orientations of the lamination, β maxima formed by flexural slip should mostly be parallel to the axes of rotation.

I. Introduction

The writer has been engaged in certain studies on structural geology in the Scandinavian Caledonides. He has made the commonplace observation that β (in this case intersections of foliation planes) of areas, sub-areas, and localities tend to form maxima of preferred orientation. These maxima do as a rule not coincide exactly with orientations of "lineations" (alignment of rock components, small folds, rods, mullions), and might, therefore, give independent clues to the tectonic history. However, orientation data, particularly β, have varying relations to the movements that gave rise to them, depending on pre-existing fabric. Thus it is frequently claimed that they may be most insecure as tectonic evidence. The purpose of the present contribution is to discuss the significance of β in superposed deformation fabrics with the aid of simple kinematic models. Lamination is referred to as S. Structures of the first deformation are called first folds. Structures of the superposed second deformation are called second folds.

II. Superposed Deformation by Flexural Slip

If the second folds are greater than, or of the same order of size as the first folds, the first folds will be rotated in the same way as ordinary lineations (Weiss, 1959, fig. 7), i.e., they will in projection travel along small circles around the axis of rotation (B) of the second deformation. In this way, the β maxima of the first folds will migrate. In cases where the second folds have rounded limbs and crests, or where they vary in intensity and enclosed angle, different parts of the first folds will be rotated to different degrees. β will thus tend to leave a trail along the projection small circle, along which the axis of the first fold travels. As for the axis of the second folds, it will not necessarily appear as a β maximum in the geometric reconstruction.

If the second folds are of a smaller order of size than the first folds,

their orientations, and thereby the orientations of β, are determined by the orientations of S of the first folds, as described by Weiss (1959). In the following list of cases that can be discerned from a kinematic point of view, it is assumed that folding by flexural slip is a way for the material to escape normal to a direction of shortening. For illustrations, see Text-figs. 1 and 2. In all cases it is assumed that the material is relatively homogeneous.

I.—Shortening in direction *a*, extension in *c*, no change of dimension in *b*.

IA.—S normal to *a*. Flattening without folding, or but slight folding with axes parallel to *b*, in connection with flow toward *c*.

IB.—S parallel to *a*.

IB1.—S normal to *c*. In the purest case, orthorhombic folding with folds parallel to *b* (Johnson, 1956, p. 348).

IB2.—S normal to *b*. Possibly no folding on any scale that gives rise to β intersections. Slight folding with axes parallel to *a*, in connection with flow toward *c*, and possibly orthorhombic wrinkles parallel to *c* (not illustrated).

IB3.—S makes angles less than 90° with *b* and *c*. Feeble folding, with axes in *bc* plane. Not of the same influence on β orientation as case IB1.

IC.—S neither normal, nor parallel, to *a*. Major folding probable only if S diverges by less than 60–70° from *a*.

IC1.—S parallel to *b*. Folding, with axes parallel to *b*. Possibly the most common case.

IC2.—S parallel to *c*. No folding, but rotation of S toward *bc* plane, possibly resulting in β orientation parallel to *c*.

IC3.—S parallel neither to *b* nor to *c*. Rotation of S toward *bc* plane, possibly with feeble folding with axes in *bc* plane.

II.—Shortening in *a*, sub-equal extension in *b* and *c*.

IIA.—S normal to *a*. Flattening without folding.

IIB.—S parallel to *a*. Folding, with axes in *bc* plane.

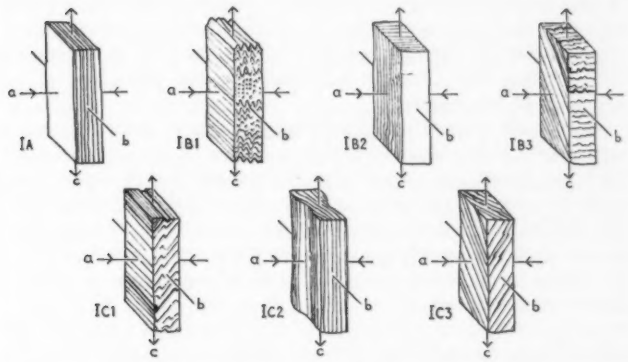
IIC.—S makes angles less than 90° with *a*. Folding, with axes in *bc* plane, except when S makes angles of more than 60–70° with *a*. In the latter case S is rotated toward the *bc* plane without any important folding.

III.—Shortening in *a* and *c*, extension in *b*.

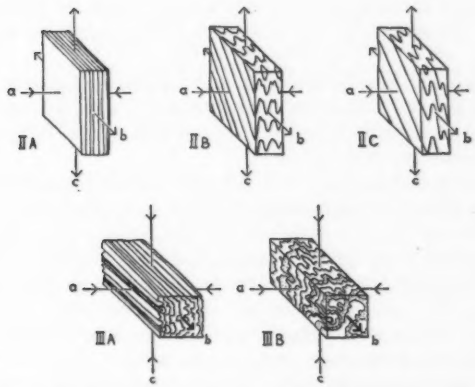
IIIA.—S parallel to *b*. Folding, with axes along *b*.

IIIB.—S not parallel to *b*. Irregular folding, leading eventually to extrusion folds parallel to *b*.

It may be objected to the cases listed that they are unrealistic, because a too passive rôle is attributed to the deformed beds. However, even when conditions permit beds to behave as competent units, they can be folded only when extension is possible normal to the potential



TEXT-FIG. 1.—Folding of surfaces with different orientations. Kinematic models with shortening along *a* and extension along *c*. The cases are commented on in the text.



TEXT-FIG. 2.—Folding of surfaces with different orientations. Kinematic models with shortening along *a* and extension along *b* and *c* (II A–C), and with shortening along *a* and *c* and extension along *b* (III A, B).

fold axis. Otherwise, the beds must yield passively. If their viscosity is too great for this, they may together constitute a unit that can yield only as a whole, and according to the " *Regel der Stauchfaltengrösse* " (Sander, 1948, fig. 50). The second folds will then no longer be smaller than the first folds (Text-fig. 3). This case was referred to at the beginning of this chapter.

The list above hardly contains anything that is novel. It should serve only as a foundation for a brief discussion on the significance of β when the second folds were formed by flexural slip. In each of the cases, the second folds occur in fields with a single orientation of S. This will be the condition of formation of the second folds when they are of a smaller order of size than the first folds. It was stated that when the change of dimension along *b* is slight or nonexistent (probably a most common case in nature), strong β maxima should develop only when S is sub-parallel to *b* (IBI, ICI). In a greater field comprising the various S orientations of the first folds, the second folds will in this case develop best, and provide the strongest β maxima, in those parts where S is most nearly parallel to *b* of the second folding. In a regional synopsis, these β maxima should dominate the pattern of β, which should thus give an approximately true indication of the orientation of *b* during the second folding. However, if the second folding meets with condition IC3 throughout a large area, it is impossible in that area to find the orientation of *b* by means of β (Weiss and McIntyre, 1957, p. 598).

If there is sub-equal extension along *b* and *c* (case II, above), β of the second folds may develop with equal strength at any orientation in the *bc* plane. This leads to the geometry shown in Weiss, 1959, fig. 6. It will be relatively rare in nature.

Shortening along *a* and *c* and extension along *b* (case III) will only lead to one stable main orientation of β, which is parallel to *b*.

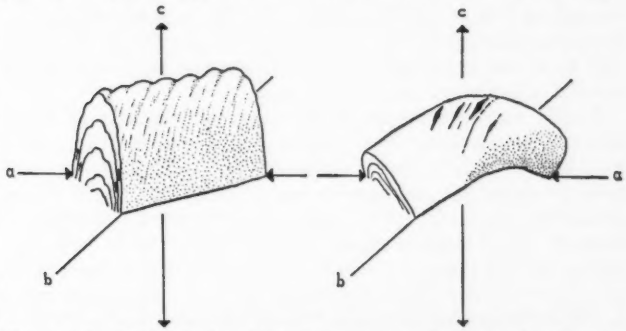
The axes of the first folds will in all the cases move along non-circular paths toward the *bc* plane of the second folding. They will ultimately approach parallelism with the direction of greatest extension. However, they are not reoriented to the same extent everywhere. In some areas that are but little affected by the second folding, the first folds will even retain their original orientation. This direction will probably be more sharply defined in a great regional synopsis than the directions of reoriented first folds that are determined by the chance amounts of reorientation in different smaller areas.

III. Superposed Folding by Non-affine Slip

Weiss (1959) has dealt with the geometry arising when second folds are formed by non-affine slip on a single set of surfaces. According to him (p. 98), β will then be concentrated along the plane of slip, at the

intersections between this plane and planes S of the first folds, and will thus give no indication as regards the direction of movement during the second folding. β of the first folding would apparently be wiped out entirely. The latter implication of the constructions of Weiss can be disproved.

By slip on a single set of surfaces, a line is rotated along a plane (in projection a great circle) toward the direction of slip (Weiss, 1959, fig. 7). A plane is rotated about its line of intersection with the slip plane (Weiss, 1959, fig. 1). This line of intersection will be the fold



TEXT-FIG. 3.—Folded beds subjected to superposed deformation in space with shortening along *a* and extension along *c*. The orientation of the beds largely prevents folding with axes parallel to the beds.
(*a*) Beds not suitably oriented for folding yield passively by flow.
(*b*) The beds cannot yield passively and cannot be folded individually, but together form a thick complex unit that is folded.

axis, with the only β concentration, when a single plane is folded by non-affine slip. However, non-affine slip acting on pre-existing first folds does not act on but one plane.

A field with non-affine slip consists of sub-fields with affine slip. It is only in such a sub-field that the orientation of any surface S can be determined as a plane. In the sub-field, all lines of a given original orientation are rotated to identical ultimate orientations. The β concentration of the first folds is thus rotated, but not dissolved, in the sub-field. In different sub-fields the β concentration of the first folds is rotated by different amounts, but always along the same great circle. For the whole field of non-affine slip, β concentrations will, therefore, be spread along the great circle through the axis of the first folding and the direction of slip of the second folding. The major β maxima will probably be near the original orientation of the first folds, for relatively undisturbed such folds, and close to the slip direction of the second folding on opposite sides of this direction, for first folds that

have been rotated in opposite senses by moderate to strong slip in different sub-fields. The presence, in the synopsis, of β with orientations that diverge from the great circle in question will depend on the extent to which S from different sub-fields are involved in the construction of β. If S from sub-fields with different amounts or senses of slip are involved in β construction, the resulting β may have any orientations, including orientations parallel to the plane of slip.

IV. Slight Flexural Folding with nearly Horizontal S

Laminated tectonites with predominantly sub-horizontal S occupy the greatest part of the Caledonian overthrust nappes in Sweden and have been dealt with in Scotland, *i. a.* by Johnson (1957, fig. 3). Rocks with a sub-horizontal mode for S provide special advantages for the use of β in the reconstruction of fold axes.

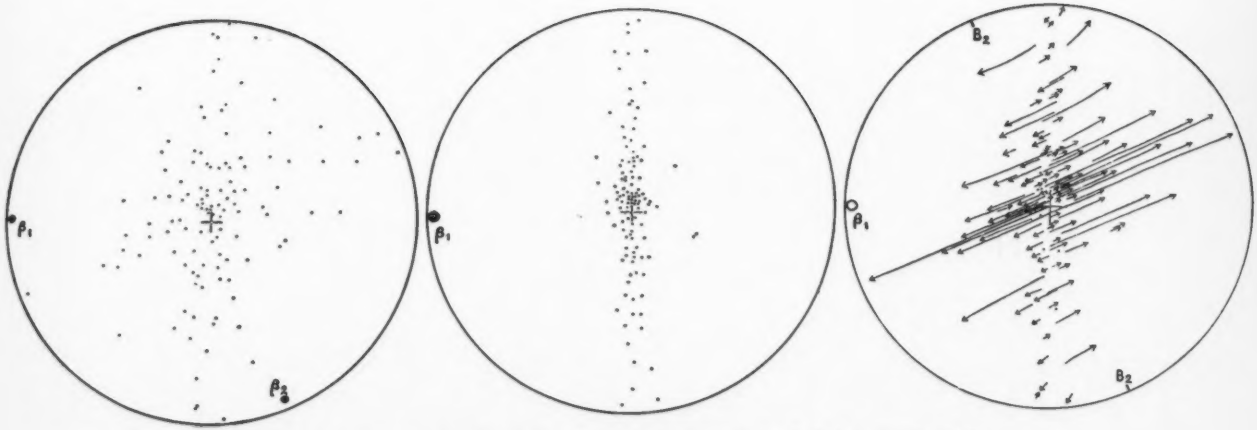
Text-fig. 4*a* shows the projection pattern of poles of S as it may appear in an investigated area. There is a sub-vertical mode that shows the S orientation to be mainly sub-horizontal. Two preferred strips of poles (π-circles) for tautozonal S can be discerned. These express themselves also in terms of β concentrations (β_1 and β_2). Text-fig. 4*b* shows an initial stage in the development of this fabric. The first folding has taken place, rotating S more strongly in some parts of the area than in others, and leaving a mode of barely rotated, sub-horizontal S. The second folding (Text.-fig. 4*c*) was of the same kind as the first, insofar as it left a mode of little rotated S. Unlike the first folding, it had to work with S that were not all horizontal at the outset. However, for any part of the pole girdle of the first folds, the modal portion of the poles were but little disturbed by the second folding. The original girdle thus remained, although considerably feebler. The sub-vertical modal portion of the first girdle supplied most poles to the reorientation during the second folding. For the second folds the densest girdle of poles thus spread out from this mode.

In areas with a single, sub-horizontal mode for S, flexural slip folding would normally have operated in the way sketched above. β maxima are in this case likely, truly to show the orientations of fold axes and axes of rotation.

V. Conclusions

In areas where complex tectonic events have resulted in multimodal, frequently steep, orientations of S, the β intersections may, as emphasized by Weiss (1959), have orientations that are random with respect to the directions of tectonic movement. However, there are certain probable and characteristic patterns for β even under these conditions.

1.—If the second folds were formed by flexural slip and are bigger than, or as big as, the first folds, β of the first folds will be spread out



TEXT-FIG. 4.—Rotation of S by two phases in field with a sub-horizontal mode for S. Points = poles of S.

(*a*) Actual orientations of S, after the deformations.

(*b*) Orientations of S, after the first deformations.

(*c*) Rotation of S by different, mostly small, amounts during the second deformation, producing orientations shown in *a*.

along small circles about the axis of the second folds. The axis of the second folds may possibly not appear as a β maximum within the space of an outcrop or an area.

2.—If the second folds were formed by flexural slip and are of a smaller order of size than the first folds, β of the second folds may in greater synopses appear mainly at right angles to the direction of greatest shortening and parallel to the direction of least dimensional change (= axis of rotation), or normal to the directions of greatest shortening and parallel to the direction of extension. β of the first folds are more or less reoriented.

3.—If the second folds were formed by non-affine slip, β of the first folds occur along the great circle through the first fold axis and the direction of slip of the second folding. If the slip was strong enough, β may be concentrated at the direction of slip.

The cases 1 and 3 can probably not be discerned by ordinary methods of counting out of β, with 1 per cent counters, but perhaps it may be possible to discern them under favourable circumstances by using very exact methods, a smaller counter, and closely spaced counts, on very great numbers of β intersections.

If the orientations of S are mostly nearly horizontal, β concentrations formed by flexural slip should reflect the regional axes of rotation in a simple way.

The discussion in this paper may be relevant, if it is assumed that the most frequent relative orientations of folds formed at the same time and under the same physical conditions are governed by principles that favour only a very limited number of such orientations. According to another, seemingly wide-spread assumption, the irregularities of tectonic processes are such that syngenetic folds may normally meet at any angle. Let the latter assumption be correct in most cases ; still, it is most infertile, chiefly because it tends to bring about neglect of the interesting means by which it could be tested.

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Origin of Albite Porphyroblasts in Rocks of the Ben More—Am-Binnein Area, Western Perthshire, Scotland

By K. A. Jones

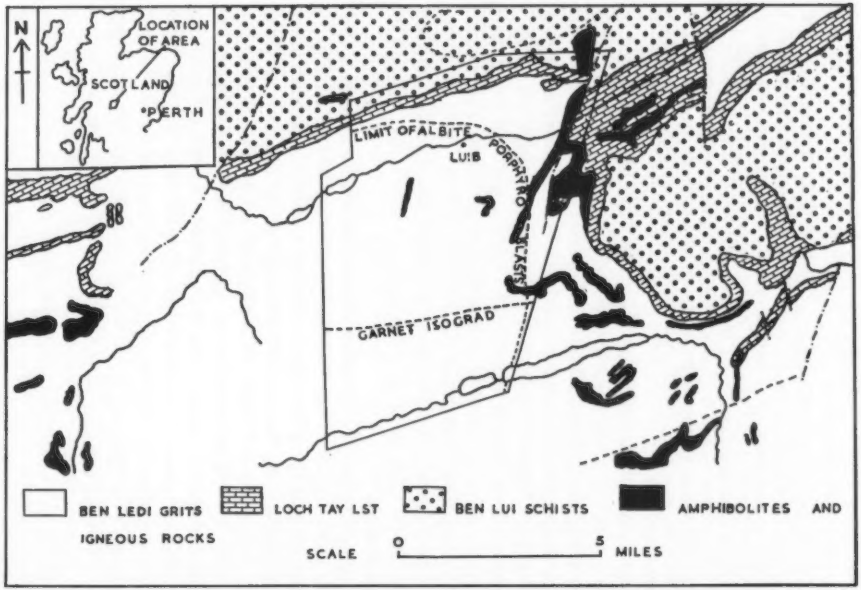
(PLATE III)

Abstract

The eastern limit of a narrow belt of rocks, running from Co. Antrim through Cowal and into Perthshire, containing abundant albite porphyroblasts, has been mapped in detail. Albite is shown to have recrystallized as large porphyroblasts (up to 23 mm. in diameter in some cases) after the last main phase of deformation affecting these rocks. It is concluded that a large scale, late stage, regional metasomatism in the metamorphic history of the rocks of this area was the causal mechanism of the growth of albite porphyroblasts.

In the present area an albite schist boundary can be drawn based on the first entry of albite porphyroblasts (Text-fig 1). The boundary between the porphyroblastic albite schists and those without abundant albite porphyroblasts continues into South West Scotland and thence into Northern Ireland, but has only been mapped in the Ben More—Am-Binnein area by the writer. Part of the work on the present problem involved an attempt at elucidating the nature and tectonic position of the limiting boundary and the origin of the albite porphyroblasts. Detailed petrographic, petrofabric and some chemical work was carried out on the rock types occurring both within, and adjacent to, the albite zone.

Clough (1897, pp. 41, 48) noticed that albite schists, associated with garnet, occur in Cowal and Cunningham-Craig (1904, p. 24, Fig. 3), drew the boundaries of four zones of progressive metamorphism in the Loch Lomond area, the highest boundary being the limit of albite gneiss. According to Clough, if the albite schists were developed from rocks with a phyllitic character an appreciable soda metasomatism must have occurred. Bailey (1923) observed that, broadly speaking, both albite and garnet are confined to separate areas, but that there was some marginal overlapping of zones and his map shows the limit of albite schists immediately to the east of Loch Lomond. He concluded that the albite schists were developed from a mixture of greywacke and shale. Cunningham-Craig (1904), Bailey and McCallien (1934) and Tilley (1925) also expressed the opinion that soda was not derived from outside the rocks in which albite porphyroblasts occur. Reynolds (1942) drew the limit of albite schists after Bailey (op. cit.) to the north-east of Loch Lomond, and stated that the albite porphyroblasts were the result of large scale soda metasomatism along the core of the hypothetical Carrick Castle recumbent anticline. Trendall (1953) has revised this isograd and the present author has modified the north-eastern extremity of the zone, by moving the boundary half a mile northwards and one mile eastwards. However, by the very nature of



TEXT-FIG. 1.—Outline map, showing the area examined in detail, and its general geological setting, drawn from observations by the writer and Sheet 38 of the Geol. Surv. Scotland.

the boundary, its limit can only be indicated approximately. The origin of the porphyroblasts was considered by Trendall to be due to a soda metasomatism, possibly from a trondjemitic magmatic source.

A detailed account of previous research in the area has been given by Reynolds (op. cit. pp. 43–45).

The Limit of Albite Schists

The albite " zone " in the area mapped is confined to the Ben Ledi Grits and its northern boundary runs parallel to the regional strike the predominant regional foliation, i.e. parallel to the strike of formations such as the Loch Tay Limestone. The nature of the boundary is slightly different inside the garnet zone north of the River Dochart, to that outside, as seen to the east of Monachyle Glen. In both cases, the boundary of the " zone " is purely arbitrary and is made by megascopic observations. Statistical microscopic work would lead to a slight extension of the " zone " and would give a more precise delineation, but it is doubtful if the extra work entailed would result in a modification of any genetic hypotheses concerning the albite in these rocks. The northern boundary of the " zone " marks the limit of the schists (mainly metapelites) containing abundant albite porphyroblasts, although outside this boundary porphyroblastic albite occurs sporadically. Along this boundary the rocks show more evidence of recrystallization than those on the eastern boundary of the albite " zone " and original sedimentary grains are consequently not so common. Immediately outside the zone, albite occurs sometimes in small isolated porphyroblasts in meta-pelites containing relatively little and often no quartz. Megascopically similar meta-pelites occur in the Kirkton Glen, but here they have no albite associated with them and contain only chlorites and light micas. Inside the albite zone these extreme meta-pelites are characteristically crowded with albite crystals, often with an equidimensional form, or occurring as short prisms after C and slightly tabular after (010). In the Ben More area the occurrence of albite reaches its maximum amount, rocks sometimes, through rarely, containing 70 per cent albite. The average size of the albite is 0·4 mm. in diameter, but grains up to ¼″ across may occur.

Petrography of the Albite Zone

The Ben Ledi Grit Group, to which the albite " zone " is confined, consists of a series of highly deformed meta-greywackes, meta-pelites and rocks transitional between these two types. It is not possible to generalize about the origin of the commonly occurring alternating bands of pelitic and more quartzitic material. Each case must be examined separately. The alternations can, in various cases, represent

TEXT-FIG. 1.—Outline map, showing the area drawn from observations by the writer and Sheet 38 of the Geol. Surv. Scotland.

original differences in lithology, subsequently modified during later deformation phases. They can also be features emphasized and in certain cases, wholly acquired by a combination of tectonic and metasomatic differentiation—a process in which shearing played a dominant role in this area. The deformation history of these rocks is too complex to be discussed here but will be described in a later publication.

Albite occurs in every lithological type in the zone, but in some specimens porphyroblasts are not abundant, though rarely are they completely absent. The rocks in the northern part of the zone have been regionally metamorphosed to the almandine grade (Text-fig. 1) and everywhere, strong chloritization of the rocks has occurred. This has often resulted in the complete destruction of garnet, with the formation of chlorite knots pseudomorphing garnet and biotite. This regional diaphthoresis is in some cases syn,— in others, post-tectonic and is, in general, synchronous with the crystallization of albite porphyroblasts. Unaltered biotite also occurs, but mainly in the more quartzofelspathic rock types.

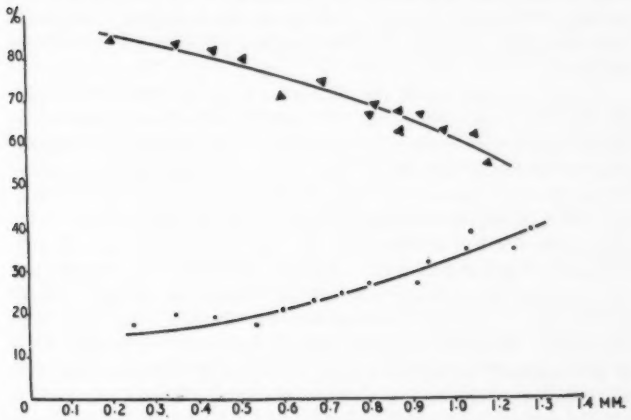
(i) *Albite schists.*—The term "albite schist" is applied to those schists containing abundant albite porphyroblasts. Porphyroblastic albite occurs typically in lenses and bands associated with the flaky minerals—light micas and chlorites. The petrography of the schists inside and outside the garnet zone is similar and all the albite schists in the area under discussion show strong evidence of recrystallization. Apart from the presence or absence of garnet (almandine), mineral assemblages remain constant throughout the area, for any given lithology. Every mineral assemblage outside the garnet zone can also occur inside it, even to the extent of absence of garnet. Albite schists typically contain quartz, chlorite and muscovite, sometimes with small and variable amounts of biotite, almandine, epidote, calcite, tourmaline, zircon, apatite, sphene, ilmenite and magnetite. Albite typically composes more than 30 per cent of the rock. This is shown in Table 1 (1,2,3,4) which are fair representative modal analyses.

TABLE 1

	1.	2.	3.	4.	5.	6.	7.	8.
Quartz .	24·32	14·08	1·10	13·40	1·37	15·32	1·87	67·41
Albite .	47·15	47·10	43·50	42·56	36·59	34·52	33·66	12·89
Muscovite	4·23	18·90	27·20	16·66	37·21	36·73	23·18	8·17
Chlorite .	1·74	13·51	25·70	27·20	22·09	8·06	28·33	3·70
Biotite .	15·56	—	—	—	—	—	—	4·32
Garnet .	—	—	0·60	0·07	—	4·67	—	0·68
Accessory Minerals	7·00	6·41	1·90	0·11	2·74	0·70	12·96	2·83

The composition of the albite is An_{4-6} and is remarkably constant. The crystals rarely show undulose extinction or any sign of post-crystalline deformation. The crystals, in the main approximately

equidimensional, vary from 0·12–20 mm. in diameter. The average size being 0·4 mm. The nature of the twinning, in those cases determined by the writer, is of the polysynthetic albite (with modifications), and in three cases only, pericline (acline) types. No C-twins have been observed. The percentage of twinned crystals varies between 17·2–38·9 (Text-fig. 2) and this figure also shows a correlation between



TEXT-FIG. 2.—Graph showing relationship between average grain size of approximately equidimensional albite porphyroblasts and percentage of grains twinned in any one specimen. (The results of analyses of fourteen specimens are given.)

▲—untwinned grains ; ●—twinned grains.

percentage of twinned crystals and average grain size of albites in the rock examined. There appears to be no correlation between An content and percentage of twinned grains. These relationships have also been observed by Gorei (1951 pp. 884–901. Figs 9 and 10). No progressive increase in twinning in the transition from the high chlorite (= diaphthoretized biotite) to the low almandine zone has been observed, although it seems to be well established that twinning in plagioclase is more frequent in rocks of moderate to high metamorphic grade, than in albite schists of the Greenschist facies. Phillips (1930 p. 225) records increasing abundance of twinned albites in passing from the chlorite to the biotite zones and Turner's (1951, pp. 581–589) observations confirm this tendency. Goldschmidt (1911, pp. 293, 295) records plentiful albite twinning in oligoclase and andesine of the Oslo hornfelses.

No satisfactory general explanation of the occurrence and amount of plagioclase twinning in felspars from igneous and metamorphic rocks has been suggested, to the author's knowledge, but several workers have expressed views on various aspects of felspar twinning (Donnay (1940, 1943) ; Emmons and Gates (1943, pp. 287–303), Turner (op. cit.)).

Turner (op. cit.) pointed out that sodic plagioclase inverts from a high to low temperature form at just below 700°C and Tuttle and Bowen (1950, p. 583) referred to the fact that plagioclase twinning is, in general, of a much more simple kind in metamorphic rocks than that in igneous rocks.

In Emmons' and Gates' (op. cit.) opinion, polysynthetic twinning is mainly a late stage phenomenon in crystal growth and is caused by mechanical strain of one sort or another. They point out that plagioclase is sometimes untwinned when set in a fine grained groundmass. When, however, the rock is mechanically deformed during crystallization, the aphanitic groundmass can exert strong enough pressures on the phenocryst for the development of twinning. Moreover, the twin frequency is stated not to be a function of composition, but is more likely to be due to properties such as viscosity or rigidity. In the majority of twinned albites examined from the S.W. Highlands there is certainly no visible external sign of significant mechanical strain at any stage of the crystal growth, even though some crystals have been rotated relative to one another (see later p. 51).

It is difficult to see on Emmons' and Gates' hypothesis, why a significantly larger number of crystals formed in igneous rocks should be twinned than those grown as a result of metamorphism and metasomatism, in which processes the crystal would presumably also be subjected to forces of a roughly similar magnitude on crystallization. That the chance of twinning in albite is much less than that in oligoclase-andesine is understandable on the French theory of twinning which explains the frequency and character of oligoclase-andesine as being due to a low index and high obliquity (Donnay 1940).

The above views were largely formed by their authors on petrological evidence. To say that the difference between metamorphic, largely untwinned, and igneous, almost invariably twinned plagioclase may reflect differences between magmatic and metamorphic crystallization, i.e. growth from a melt as opposed to growth by diffusion in the solid state, is merely a tautology and not an explanation. The question remains, what are the fundamental differences between magmatic and metasomatic crystallization, which possibly have a bearing on the development of plagioclase twins ?

Billig (1956) advanced the view that crystals will twin if their growth direction becomes closer to the main temperature gradient. This has been observed experimentally. It was also established by Billig, that

impurities can produce twinning. This was induced experimentally by passing NH_3 gas through the melt during silicon crystallization. The reason suggested is that there is more room on a twin boundary in the diamond lattice and the accumulation of impurities in this region is therefore accommodated by the formation of a twin. A third possible cause of twinning is by the progressive pile up of dislocations of one sort or another in the crystal lattice. Stacking faults can also give rise to twins.

Intense thermal stress has also been shown, experimentally, to lead to twin development in quartz, probably because it causes imperfections in the seed lattice.

Lawson and Nielson (1958) remarked that " To avoid twinning, one is forced once again to conclude that slow regular growth and maximum purity may be necessary and that, in particular, thermal stresses should be avoided ".

These conditions are clearly more likely to be satisfied by a relatively slow crystallization involving some slow diffusional process as, for example, in many metamorphic rocks in which albite porphyroblasts have grown than in relatively rapidly cooled igneous melts.

It seems likely to the writer, therefore, that pure " mechanical " twins are a comparatively rare phenomenon in igneous rocks. Given the above-mentioned physical conditions the French theory of twinning predicts that some felspars are more likely to be twinned in any particular way than others and felspars more than other common rock forming minerals.

The statistical correlation between average grain size of albite and percentage of grains twinned could, in certain cases, be caused by a greater susceptibility of larger grains to small differential movements within a crystalline aggregate, or to a more rapid growth from a seed lattice with imperfections of one kind or another. It is also possible that small compositional differences are significant but since the composition of the albite is, as far as can be determined with the Universal stage and R.I. measurements, constant for all grain sizes, it is unlikely that this is a controlling factor in twin growths.

The second explanation is considered the more likely to apply to the majority of albites in the rocks under discussion since there is a complete lack of evidence of the majority of these crystals ever having been subjected to any significant mechanical strain.

This hypothesis does not demand that all large crystals, even adjacent ones, should be twinned and all small ones untwinned, since many metasomatic and dynamic conditions (availability of material, channels along which solutions and " dry " ions can migrate for example) can be critical in determining whether any single crystal is large, small, twinned or untwinned. However, on a statistical basis, it is to be

expected that if the above hypothesis is correct, the relatively large grains will be more often twinned than the small grains.

Some of the albites show a simple inverse zoning, the inner zone being composed of An_{3-6} and the outer zone of An_{6-8}. All of the outer zones measured exhibit the same optical relationship to the core crystal, the β direction of the outer zone is typically 5° from β and γ is 2·5° from γ of the core crystal. The development of zoning bears no apparent relationship to lithology and since the zoning is of such a simple type it was impossible to plot, for example, An content against distance from the core. In the case of more complex crystals from higher zones than the present rocks, such measurements could be useful. Only very rarely can complete albites of An_{6-8} composition be observed in these rocks and whether this represents a new and slightly later nucleation is unknown. Attention to this inverse zoning has been drawn by Becke (1913), Carstens (1924), Phillips (1930) and Grubenmann and Niggli (1924), who suggested that since anorthite is more soluble that albite, the latter mineral first crystallizes from aqueous solution, resulting in an increase in anorthite outwards.

Invariably, at least two generations of muscovite are present, the later of which occurs as large crystals and is often developed on a considerable scale. The earlier generation is, in the main, pre-albite porphyroblast growth and the latter synchronous with it.

Inclusions in albite are typically, but not always, abundant and are mainly of quartz, epidote, magnetite, ilmenite and, more rarely, actinolite, garnet, chorite, sphene, and apatite. Inclusions of the phyllosilicate minerals, with the exception of marginal chlorite, are rare.

(ii) *Petrography of Associated Rocks.*—These fall into three broad categories (a) meta-greywackes, (b) meta-quartzites, (c) epidiorites, of which type (a) is most common. Mineral parageneses in the first two categories are similar to those in albite schists described above, but the proportions are different (Table 1, 6, 7, 8,). In meta-greywackes and -quartzites, albite occurs as (i) small, original sedimentary grains, with strongly undulose extinction (ii) slightly larger, partially recrystallized grains, often with a partial sedimentary outline, and (iii) large porphyroblasts. Sedimentary albites are best seen in meta-greywackes in which the oldest structures have been preserved . A typical analysis of a more quartzitic rock type is given in Table 3. 3.

The composition of epidiorites is more significant than that of meta-greywackes from the point of view of albite porphyroblast genesis. Geikie (1891, p. 76) first demonstrated that the Highland epidiorites have compositions similar to dolerites or basalts and Wiseman (1934, p. 372) quoted analyses, showing the remarkably close correspondence between the chemical composition of a chlorite-clinozoisite-albite amphibolite from South Knapdale and an olivine

basalt from Hawaii. The petrography of epidiorites similar (in most respects) to those in the present area has been thoroughly described by Wiseman (op. cit.) and it is therefore unneccesary to give a detailed account of the mineral parageneses but the amount and areal distribution of albite in these rocks is atypical of the standard epidiorite.

Epidiorites inside the albite zone contain an abnormal amount of modal albite, sometimes as much as 30 per cent, compared to those outside the zone, which usually contain approximately 10 per cent. This is shown in Table 2.

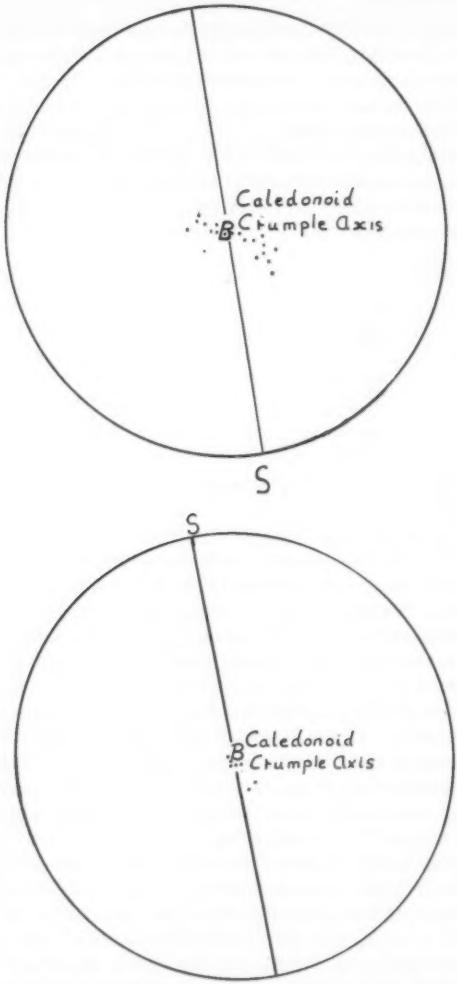
TABLE 2

	1.	2.
Actinolite . .	3·7	78·3
Albite . .	31·7	10·3
Chlorite . .	37·8	0·9
Sphene . .	10·6	5·1
Epidote . .	7·1	1·5
Calcite . .	—	3·6
Garnet . .	—	0·3
Quartz . .	9·1	—

Typical modal analyses of epidiorites inside (1) and outside (2) the albite " zone ".

Time of Crystallization.—It is probable that albite continued to grow from the earliest to the latest episode in the mineralogical development of these rocks, by processes associated with metamorphic differentiation involving sedimentary albite, but one main phase of albite porphyroblast crystallization can be distinguished. The oldest grains are the above-mentioned small, rounded, sedimentary grains occurring in largest quantity in rocks which have been least deformed. The partially recrystallized sedimentary grains possibly grew during diagenesis of the original sediment. Large porphyroblasts often contain measurable quartz inclusions and helicitic folds and thus their time of crystallization, in relation to the various phases of movement can be determined in these cases. Much care must, however, be taken with such analyses because, in certain cases, extremely misleading fabric diagrams can result and interpretations of such diagrams are invalid unless several controls are exercised (Jones 1959). Axes of small folds composed largely of magnetite, ilmenite and epidote, when plotted on stereograms (Text-fig. 3, a, b; and Plate III A and B), shows that the main episode of albite crystallization was in some cases synchronous with, and in others later than the last deformation phase affecting the rocks, a deformation which produced a strong regional crumpling of the regional foliation plane, the crenulation axes plunging statistically at 10° in a 70° direction. The detailed tectonic history of these rocks will be discussed in a subsequent communication.

The orientations of essentially planar helicitic structures in adjoining



TEXT-FIG. 3.—(*a*) Typical stereogram, showing axes of folds preserved helicitically in ten albite porphyroblasts in a single section. The centre of the stereogram is the projection of the externally measured Caledonoid crumple axis.

(*b*) A similar example to that of (*a*), but nineteen helicitic folds were measured on this section and the scatter is slightly, but not significantly, different.

porphyroblasts are sometimes mutually perpendicular and bear no relationship to the principal foliation plane, or any other plane, external to the crystal. It is probable that these structures represent the principal foliation plane in the rock in most cases (they can also represent strain-slip cleavage in some cases, but these can usually be recognized as such), and thus the crystals must have been rotated. A relatively simple mechanical rotation of the grains and their inclusions relative to each other cannot form such structures although some sort of grain rotation has obviously occurred. The explanation is probably that the grains have moved relative to one another by the removal by reaction of muscovite, chlorite and quartz mainly, thus allowing the albites to be rotated, largely haphazardly, and in certain cases, pressed together, in which position annealing crystallization has occurred.

Discussion

There are, broadly speaking, three possibilities for the origin of albite porphyroblasts in these rocks:—

(*a*) the necessary sodium is contained in the pelites;
(*b*) sodium came from felspathic greywackes and was metasomatically transferred to the pelites during metamorphism ;
(c) sodium was introduced from an extraneous source by a large scale metasomatism.

(*a*) If this hypothesis is true, the sodium must be contained in micas, from which albite porphyroblasts can be seen to grow. Thus the rocks must contain a considerable proportion of a sodic mica such as paragonite. This mineral is known to occur in fairly large amounts in the Zillertaler Schists, Austria, but it seems doubtful if it could occur in large enough quantity to form the south-west Highlands albite schists.

(*b*) This hypothesis is suggested by the abundance of clastic grains in the Ben Ledi Grits both inside and outside the albite " zone ", and by the confinement of the albite schists to the Ben Ledi Grit Group, both in the present area and in the rest of the South West Highlands and Co. Mayo (Trendall, op. cit.). Reynolds' (op. cit. p. 45) assertion that " no soda-rich examples of unmetamorphosed geosynclinal sediments are known " is not consistent with soda percentages in greywackes cited by Pettijohn (1948, p. 250). These percentages, 2·68, 3·78, 4·21, 2·50, 3·13, make it quite clear that sediments containing enough sodium to give, on metamorphism, albite schists, certainly exist.

However, albite porphyroblasts can be seen growing in meta-greywackes and adjacent meta-pelites simultaneously and there is every indication that no metasomatic transfer of soda from the greywackes to the pelites has occurred. Meta-greywackes inside the albite " zone,"

for example, contain the same amount of felspar as those outside the " zone ".

(*c*) The third hypothesis is held, by the writer, to be the most probable explanation of the origin of albite porphyroblasts. This broad hypothesis has also been advanced by Reynolds (op. cit.) and Trendall (op. cit.) but the present writer's hypothesis has been formulated using evidence which departs, in detail, from those of the above writers.

Reynolds' evidence for metasomatism was obtained from several chemical analyses of albite schists and non-albitic rock types in the immediate vicinity of each other. Two analysed specimens, one containing a moderate amount of albite and the other practically albite-free are described by Reynolds as occurring in the same horizon within two feet of one another (Table 3, 1 and 2).

TABLE 3

	1.	2.	3.
SiO_2	40·31	51·92	81·55
Al_2O_3	22·52	20·47	9·73
Fe_2O_3	2·31	1·51	0·54
FeO	11·49	6·98	0·82
MgO	6·80	4·67	1·15
CaO	1·85	1·19	0·99
Na_2O	2·84	4·16	2·78
K_2O	6·87	4·70	0·97
H_2O+	2·99	2·70	0·66
H_2O-	0·20	0·18	—
CO_2	—	tr	0·70
TiO_2	1·13	0·63	0·01
P_2O_5	0·11	0·44	—
S	0·01	tr	—
MnO	0·11	0·10	—
BaO	0·35	0·16	—
SrO	—	—	—
	99·89	99·89	100·00

1. Biotite schist, with some albite porphyroblasts, from Ben Ledi Beds, Portna Lorna, Co. Antrim. (Reynolds, 1942 ; analyst, Agnes Gibbs.)
2. Albite schist, from the same horizon as specimen 1 and collected less than two feet from 1. (Reynolds, 1942 ; analyst, Agnes Gibbs.)
3. Meta-greywacke, Monachyle Glen. Analyst, T. W. Bloxam.

According to Reynolds, these analyses show that sodium silicate and P_2O_5 have migrated into the biotite schist, thus forming an albite schist, and potassium, iron and magnesium have been driven out. Further, it is claimed that neither the biotite nor albite schists have the compositions of altered sediments and that the low silicon, combined with high alkali, iron and magnesium of the biotite schist make it " . . . clear that there was an introduction of at least potash before the introduction of sodium silicate which led to the development of albite

porphyroblasts". However, if the tectonic and metamorphic history of the Ben Ledi Grits is taken into account, the composition of the analysed biotite schists could not be expected to be comparable to that of any geosynclinal sediment. The complex geometry of the structures in the Ben Ledi Grits of the area under discussion has been briefly described by Bowes and Jones (1958). As a result of this tectonic style shown by the rocks, the main schistosity in most cases replaces and is not coincident with the old bedding planes. However, on a very small scale, there is often no certainty that two specimens collected even 1″ from one another along the strike of the main regional foliation planes were once part of the same sedimentary horizon. Many examples, clearly indicating the need for great caution when trying to correlate bands, even over a distance of inches, can be seen in the Ben More—Am-Binnein area. It is, however, possible that Reynolds was aware of these difficulties although no mention is made of it in the paper.

Further evidence of chemical change given by Reynolds is that, " . . . it is evident from thin sections of the albite schists that migration of material has taken place. The change in colour of the biotite from brown to bluish-green as the albite porphyroblasts develop is evidence of chemical change." Detailed work by the writer has, however, established that green biotite is not the most common type in the region investigated, the brown variety being more typical. When green biotite occurs in these rocks it is possible that it represents a stage in the diaphtoretic change of biotite to chlorite (Winchell, 1951, p. 376).

One further point concerning Reynolds' hypothesis is that the soda metasomatism occurred along the core of the Carrick Castle fold. It is not clearly stated, however, where the root-zone of this fold and thus, presumably, the source of the soda is located and the sections published in the 1942 paper do not go far enough north for one to be able to deduce the precise location of this zone. Moreover, detailed fabric work on a macro- and microscopic scale makes it clear that if a gigantic recumbent fold exists it must have been formed earlier than any of the three main deformation phases recognized in the area, and its effect on the rock fabric must have been totally obliterated. This is considered to be extremely unlikely. Since the albite porphyroblast crystallization is almost the latest event in the crystallizational history of the area, a considerable time gap must also have existed between the formation of the Carrick Castle fold and soda enrichment along its core.

Trendall's hypothesis that the albite schists were formed by a metasomatism associated with a " late para-tectonic phase of metamorphism to some extent unconnected with the earlier regional metamorphism " has much in common with Reynolds' hypothesis, in that both hypotheses agree that large scale soda metasomatism has occurred.

All the evidence of the present writer, however, suggests that the

formation of albite schists is an integral part of the regional metamorphism of the area. It is significant that albite never contains large quantities of muscovite although quartz, magnetite and epidote and other minerals often occur in abundance. The textural relationship between albite porphyroblasts and the flaky minerals shows convincingly that albite has used muscovite in its formation. For example, albite has often grown in micaceous bands by absorbing muscovite in intergrowths of muscovite and chlorite, leaving part of the chlorite as marginal inclusions. To explain the amount of albite present, however, a large scale influx of sodium must be invoked, since analyses suggest that not enough sodium could have existed in the original rocks.

Albite is often later than the development of garnet, but can also be contemporaneous, and is mainly associated with the retrograde metamorphism of the area. Thus no time gap exists between regional metamorphism of the area and the main period of albite crystallization. It is suggested that albite continued to grow from the earliest to the latest stages in the mineralogical development of the rocks of this area. Some of it certainly came from original sedimentary grains. However, it is suggested that most of it arose as the result of a late regional soda metasomatism.

ACKNOWLEDGMENTS

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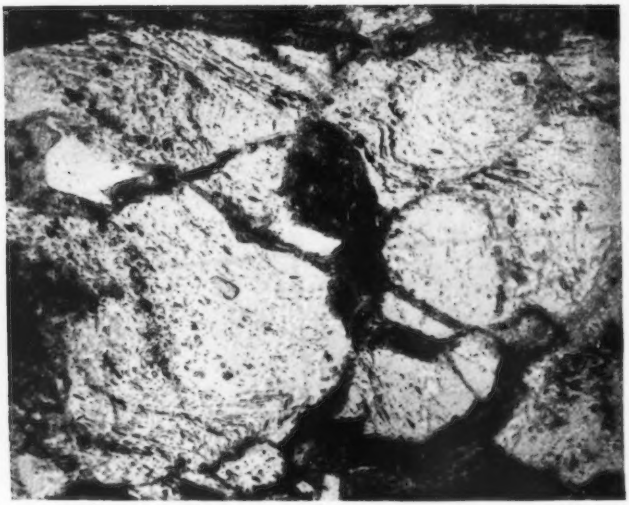

B

STRUCTURES IN ALBITE PORPHYROBLASTS, WEST PERTHSHIRE, SCOTLAND.

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EXPLANATION OF PLATE III

A.—Photomicrograph of helicitically preserved microfold (part of a strain-slip structure) in albite porphyroblast. (Ordinary light × 90.)

B.—Photomicrograph of helicitic structures in albite porphyroblasts (Ordinary light × 100.) (Parts of the photomicrograph are slightly out of focus due to tilting of the thin section on the U-stage.)

Hydrothermal Melting of Shales*

By P. J. Wyllie† and O. F. Tuttle

Abstract

PT curves for the beginning of melting of five analysed shales in the presence of water vapour under pressure are 20° C. to 40° C. higher than the corresponding curve for granite. About 150° C. above the beginning of melting, the shales are half-melted ; this is higher than the liquidus curve of most granites. Refractive indices of the quenched liquids (1·495–1·505) indicate a granitic or granodioritic composition. Quartz, cordierite, mullite, hypersthene, anorthite, etc., are developed in the partially fused shales. Partial fusion of shales by a granitic magma, even if superheated, would produce a liquid no more basic than granodiorite. The chemical characteristics of the shales are compared with average igneous rocks, and there appears to be no possibility that fusion of shales could produce a basaltic magma. Complete fusion would produce a melt with composition distinct from normal igneous magmas.

Introduction

It has been suggested that basic igneous magmas could be produced by the fusion of shales (cf. Bowen and Aurousseau, 1923). Although it is now established that granitic magmas can be obtained by the fusion of feldspathic sandstones or by the partial fusion of other rocks (Knopf, 1938 ; Wyllie and Tuttle, 1960b), the development of more basic magmas by the fusion of sediments is complicated by thermal and compositional difficulties. Bowen and Aurousseau (1923) critically examined the possibility that normal igneous magmas could be derived by the fusion of sediments, and they pointed out that despite the broad similarities in composition between the average igneous rock and the average shale, there are distinct chemical differences which are thoroughly characteristic. The average shale contains a marked dominance of K_2O over Na_2O, and an excess of Al_2O_3 over that required to make aluminosilicates with alkali and lime. Liquids produced by the fusion of shales should therefore contain such minerals as sillimanite and cordierite. These minerals are characteristic of fused shale xenoliths, but not of uncontaminated basaltic rocks.

Oftedahl (1957) has suggested that the composite dykes of the Oslo region, granitic sills with diabase margins, were produced by the fusion of shaly sediments in contact with granitic plutons. He envisaged the local development of basaltic magmas at the contacts, and assumed that when the granite was intruded to higher levels as dykes, it was preceded by the less viscous basaltic magma which thus formed the diabase margins. Whenever the origin of composite intrusions is

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investigated the conclusion seems inevitable that contrasting magma types coexist at depth (e.g. Rao, 1958). However, the hypothesis that basaltic magma could be produced by the fusion of shales is unlikely in the light of existing evidence (Bowen and Aurousseau, 1923), and the experimental results described below confirm this. The product when shales are partially fused is a granodioritic liquid containing aluminous minerals.

Method

Five analysed "standard shale samples" from Pennsylvania, varying considerably in chemical composition (Table 2), were kindly provided by Dr. M. L. Keith. All samples contain 5 to 10 per cent of volatile constituents.

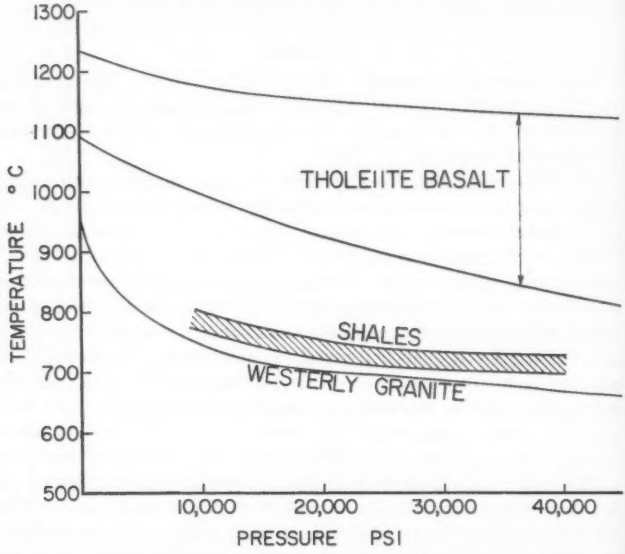
The shales were sealed within gold capsules, together with a little water, using the technique described by Tuttle and England (1955). The small closed systems were placed within cold-seal pressure vessels (Tuttle, 1949), pressure was applied by water, and the temperature was raised to the required level. The equipment used was essentially that described in detail by Harker and Tuttle (1955). Pressure was transmitted to the samples by collapse of the capsule walls. At the end of a run, the vessel was quenched rapidly in cold water. The sample was removed from the capsule, crushed, and examined by X-ray powder diffraction methods, and with a petrographic microscope.

Results

When our results were presented at the Washington Meeting of the American Geophysical Union in May, 1958 (Wyllie and Tuttle, 1958), Winkler (1957) had already published some studies on the experimental metamorphism of lime-free clays at 2,000 atmospheres pressure. Since then he and von Platen have published additional studies on lime-bearing clays (1960), and on the effect of NaCl as well as water (1958). He was concerned mainly with the stability of mineral phases developed in the clays at sub-solidus temperatures, and found that runs of 14 days' duration were required for the attainment of equilibrium. Melting began at 700° C. at this pressure, and the calculated composition of the liquid at 740° C. corresponds to that of a leuco-granite. Addition of NaCl lowered the melting temperature to 670° C.

Our runs were aimed at determining the PT curves for the beginning of melting of the five shales for comparison with similar curves for other rocks. The results are illustrated in Text-fig .1, and a selection of the completed runs are listed in Table 1. The first three shales were studied in fair detail, but few runs were made using the other two samples. None of our runs lasted longer than 3½ days, but we believe this to be sufficient for recognition of the beginning of melting. At

higher temperatures, when the amount of liquid increased, equilibrium is more readily attained than in sub-solidus runs. The minerals which were identified either from X-ray patterns or with the microscope are listed in Table I, but no attempt was made to study the mineral phases in detail. It is unlikely that the minerals listed represent equilibrium assemblages for runs containing little or no liquid.



TEXT-FIG. 1.—Measured melting temperatures of rocks at various water vapour pressures. The shaded band represents the range of beginning of melting of the five analysed shales listed in Table 2. The beginning of melting of Westerly granite (Tuttle and Bowen, 1958) and both solidus and liquidus of a tholeiite basalt (Yoder and Tilley, 1956) are given for comparison.

Abundant cordierite developed in many runs. Its properties were similar to those of the microscopic cordierites found in fused shale and sandstone xenoliths. Sector twinning occurred frequently, and in most charges containing cordierite a few crystals were found with cruciform twinning. This type of twinning in cordierite was described recently in fused xenoliths of Torridonian sandstone (Wyllie, 1959).

Some fused charges containing cordierite were cooled slowly to sub-solidus temperatures. This produced pinite—the cordierite was progressively replaced from its margins by muscovite. Anorthite developed in the shale which contained calcite (Tables 1 and 2).

TABLE 1.—RESULTS OF HYDROTHERMAL EXPERIMENTS ON SHALES LISTED IN TABLE 2

Abbreviations: L—liquid; Q—quartz; Ml—mullite; Mv—muscovite; C—cordierite; A—anorthite; H—hypersthene; (t)—trace; psi—pounds per square inch.

Shale	*Pressures* (psi)	*Temp.* (° C.)	*Time* (hrs.)	*Result*
56–42–1	10,000	780	27	Q, Ml.
,,	,,	800	90	Q, Ml, L (t).
,,	,,	940	23	Q, Ml, L.
,,	20,000	730	48	Q, Ml.
,,	,,	750	24	Q, Ml, L (t).
,,	40,000	710	46	Q, Ml, Mv.
,,	,,	725	46	Q, Ml, Mv, L (t).
,,	,,	820	4	Q, Ml, L.
56–43–1	10,000	770	13	Q, An, Mv.
,,	,,	780	27	Q, An, Mv, L (t).
,,	,,	940	22	An, H, L.
,,	20,000	730	48	Q, An, Mv.
,,	,,	750	24	Q, An, Mv, L (t).
,,	40,000	710	46	Q, Mv, An, H (t).
,,	,,	725	46	Q, Mv, An, H (t), L (t).
,,	,,	820	4	An, Q, Mv, H (t), L.
56–309–1	10,000	770	13	Q, Mv, Ml.
,,	,,	780	27	Q, Mv, Ml, L (t).
,,	,,	940	46	C, L.
,,	20,000	720	80	Q, Mv, Ml (t).
,,	,,	730	48	Q, Mv, Ml (t), L (t).
,,	40,000	710	46	Q, Mv, Ml, C.
,,	,,	725	46	Q, Mv, Ml, C, L (t).
,,	,,	780	3	Q, Mv, Ml, C, L.
56–316–1	10,000	830	26	Q, C, Ml, L.
,,	,,	940	23	C, Ml, L.
,,	40,000	720	23	Q, Mv, Ml.
,,	,,	735	25	Q, Mv, Ml, L (t).
56–462	10,000	940	23	Ml, L.
,,	40,000	710	42	Q, Mv, Ml.
,,	,,	720	23	Q, Mv, Ml, L (t).
,,	,,	820	4	Ml, Mv, L.

At a given pressure of water vapour, all five shales begin to melt in a temperature interval of about 20° C. The shales are very fine-grained, and the temperature differences are probably due to the difficulty of detecting the first trace of liquid in such fine-grained assemblages rather than to compositional differences among the shales. The shaded band in Text-fig. 1, shows the measured PT conditions for the beginning of melting of the shales. The melting temperature at 2,000 atmospheres (29,000 p.s.i.) is close to the value obtained by Winkler (1957) for lime-free clay. The band is about 20° C. higher than the PT curve for

the beginning of melting of granite in the presence of water vapour (Tuttle and Bowen, 1958). The PT curves for the solidus* and liquidus of tholeiite basalt in the presence of water vapour (Yoder and Tilley, 1956) are given for comparison.

Shales begin to melt at temperatures not greatly in excess of the melting temperatures of granite. The actual difference is probably

TABLE 2.—CHEMICAL ANALYSES OF "STANDARD SHALE SAMPLES"

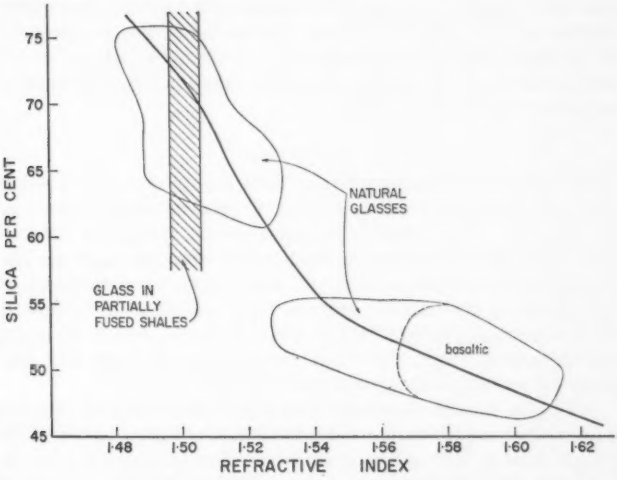
Samples supplied by Dr. M. L. Keith, College of Mineral Industries.

Sample designation :	56–42–1	56–43–1	56–309–1	56–316–1	56–462
SiO_2	59·2	54·1	52·8	60·9	37·7
TiO_2	1·56	0·89	0·94	1·04	1·04
Al_2O_3	24·4	18·1	19·0	21·2	26·9
Fe_2O_3 (total iron)	2·63	4·54	7·84	3·94	10·38
MnO	0·02	0·08	0·18	0·04	0·12
CaO	0·44	6·93	1·80	0·22	0·10
MgO	0·39	1·30	1·64	1·73	1·44
K_2O	1·81	2·87	3·20	3·60	4·38
Na_2O	0·14	0·12	0·21	0·15	0·24
Li_2O	0·031	0·012	0·035	0·015	0·043
BaO	0·27	0·10	0·08	0·08	0·12
P_2O_5	0·34	0·28	0·25	0·10	0·14
F	0·08	0·07	0·07	0·07	< 0·01
S	1·00	0·04	2·40	0·26	5·10
CO_2	0·10	5·08	1·65	< 0·05	< 0·05
C	0·72	0·29	3·31	1·31	1·01
$H_2O > 110°$ C	7·0	4·4	5·0	5·0	7·34
H	0·14	0·14	0·28	0·22	7·0
$H_2O < 110°$ C	0·90	1·40	1·00	1·02	0·41
	101·17	100·74	101·68	100·90	103·46
Less O for S	0·38	0·01	0·90	0·10	1·91
Less O for F	0·03	0·03	0·03	0·03	—
	0·41	0·04	0·93	0·13	1·91
Total sum	100·76	100·70	100·75	100·77	101·55

less than that measured because of the fine-grained nature of the shales. The melting interval of most granites is quite small, so the PT curve for a granite liquidus would not be far above the temperature of beginning of melting of the shales, and certainly below the solidus for the tholeiite, which is 150° C. to 250° C. higher than the granite solidus. Most granites would thus have to be superheated to cause any melting in crystalline basalt, and a prohibitive amount of superheat would be required for the complete melting of basalt by a granite magma.

* The term solidus is used here to indicate the lowest temperature at which microscopic evidence was found for the existence of a liquid phase. In these complex systems it is likely that a small amount of liquid may develop at considerably lower temperatures.

The first liquid developed by fusion of the shales is granitic or granodioritic in composition, as shown by the refractive index of the glass in quenched charges. 150° C. above the temperature of beginning of melting, the fused shales consist of at least 50 per cent liquid, together with crystals of quartz, mullite, cordierite, hypersthene, etc. These conditions correspond approximately to the beginning of melting of basalt, but the liquid present is still granodioritic in composition. Much higher temperatures would be required to melt the shales completely.



TEXT-FIG. 2.—Curve showing the relationship between the silica contents and refractive indices of natural glasses (George, 1924). The vertical shaded band shows the range of refractive indices measured for quenched liquids in the experimentally fused shales.

Text-fig. 2, shows George's (1924) curve for the refractive indices of natural glasses plotted against their SiO_2 percentages. This curve gives a rough approximation of the SiO_2 content of a glass from its refractive index. The two areas marked in Text-fig. 2 enclose all the points plotted by George. Acid and intermediate glasses have low refractive indices, and basic glasses have much higher indices.

Refractive indices of the glass in each charge were measured and, for all five shales, from lowest to highest temperatures, they were between 1·495 and 1·505. The range of values is shown in Text-fig. 2 as a vertical shaded band. According to George's curve, the glasses should contain between 70 per cent and 73 per cent SiO_2. We are not

altogether satisfied with the usefulness of this curve as an absolute measure of SiO_2 content (Tilley, 1922), but the measured indices make it clear that the liquid developed is acid, and far removed from basaltic compositions, even when the shales are half-melted.

Because the shales themselves contain such high proportions of volatiles, several runs were made without the addition of water. These appeared to produce the same results as those with additional water, but the reactions were more sluggish.

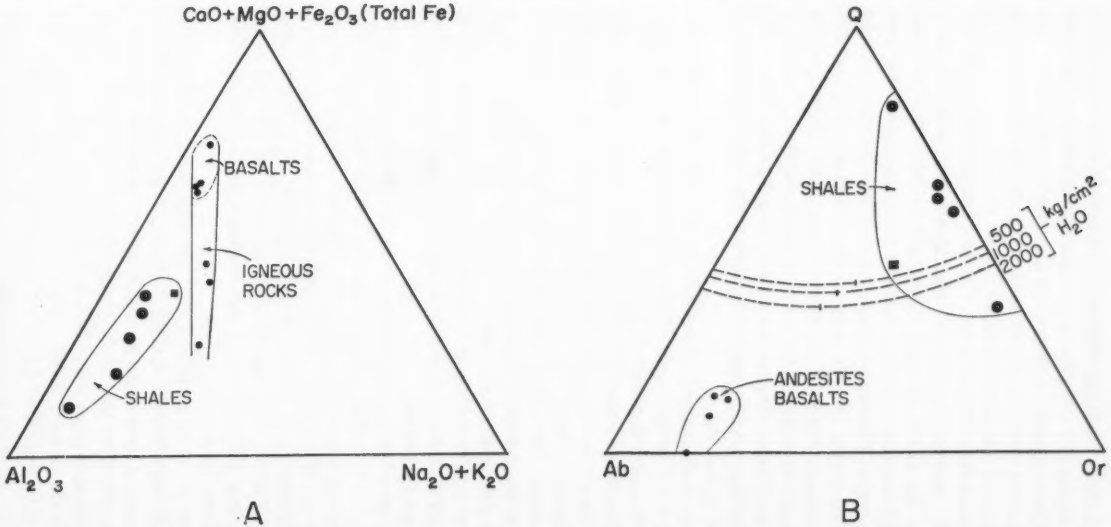
The half-melted charges at high temperatures were extremely vesicular. Granite liquid under similar conditions contains many small bubbles when quenched to a glass, but the frothy, slaggy appearance of the fused shales is quite different. Continued release of volatile materials from combination in crystalline phases is a possible cause of the large and abundant vesicles.

Discussion

Experimental results demonstrate quite clearly that an impossible amount of superheat would be required to produce basaltic magma by the fusion of shales at granite contacts (Text-fig. 1), and the compositional difficulty is illustrated by the fact that when the shales are half melted, the liquid produced is still acid, probably approaching in composition a granodioritic magma. Any superheat in magmas is likely to be small (Bowen, 1928, p. 184). However, even if the shales were completely fused, their bulk compositions are such that they would not resemble any normal igneous magma.

In Text-fig. 3A, the shales have been plotted on a triangular diagram in terms of alumina, alkalies, and total cafemic oxides. Clarke's (1924) average shale is plotted for comparison. This is relatively lower in Al_2O_3 than the Pennsylvania shales. The compositional trend of igneous rocks is illustrated by some of Nockolds' (1954) averages, ranging in composition from basalt to granite. The shales have an excess of Al_2O_3 compared to the igneous rocks, and their bulk compositions are more closely related to acid and intermediate rocks than to basalts. In both experimentally and naturally fused shales, excess Al_2O_3 causes the crystallization of aluminous phases such as cordierite, sillimanite, mullite, etc.,

Text-fig. 3B is a projection on to the anhydrous base of the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O of the isobaric quartz-feldspar boundaries for water vapour pressures of 500, 1,000 and 2,000 kg./cm.^{2} (Tuttle and Bowen, 1958). The boundaries separate the primary crystallization fields of feldspar solid solution and a silica phase. The temperature minimum on each boundary is marked by a short cross-line. The difference in the alkali ratios of basalts and the shales is shown by plotting their respective bulk compositions (recalculated to



TEXT-FIG. 3.—The chemical composition of shales compared with that of some igneous rocks. Large circles—shales listed in Table 2 ; small circles—Nockolds' (1954) average igneous rocks ; small square—Clarke's (1924) average shale.

A. Comparison in terms of alumina, alkalies, and cafemic oxides, recalculated to 100 per cent.

B. Comparison in terms of normative albite, orthoclase, and quartz recalculated to 100 per cent. The dotted lines are projections of isobaric quartz-feldspar boundaries in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O. The compositions of granites are grouped near the minimum temperatures (cross-lines) on these boundaries.

100 per cent after omitting volatiles) in terms of normative albite, orthoclase, and quartz. Volatiles were omitted, because if a shale were completely fused it would lose most of its volatile components. The shale norms were calculated in the usual way so that they could be compared directly with igneous rocks.

The five shales from Pennsylvania, Clarke's average shale, and some of Nockolds' average basalts and andesites have been plotted in Text-fig. 3B. The shales are much richer in K_2O/Na_2O than the basic igneous rocks, and the Pennsylvania shales are richer than the average shale in this ratio. The basic rocks occupy an area far removed from the shales, and on the opposite side of the projected isobaric quartz-feldspar boundaries (with the exception of one shale). When granites are plotted on this diagram, they are concentrated near the minima on the quartz-feldspar boundaries (Tuttle and Bowen, 1958). The shales are thus more closely related to acid than to basic igneous rocks when plotted in this way.

Further compositional differences are illustrated by the Differentiation Indices (D.I.) of the rocks. The D.I. of a rock is defined as its percentage of normative salic minerals, quartz, orthoclase, albite, nepheline, leucite and kaliophilite (Thornton and Tuttle, 1956). None of the rocks under consideration contains nepheline, leucite or kaliophilite in the norm. Therefore, if we take a tetrahedron with the base shown in Text-fig. 3B, and the apex representing all other normative minerals, then the D.I. gives the percentage distances of each rock from the apex of the tetrahedron towards its base. The D.I.'s for basic rocks are low, and those for acid rocks are high. The points plotted in Text-fig. 3B give the projection of each rock from the apex of the tetrahedron on to its base.

The D.I.'s of the average basalts and andesites plotted in Text-fig. 3B range from 17 to 43 ; they lie in the upper half of the tetrahedron. The D.I.'s of granites range from 80 to 100 ; they lie close to the base of the tetrahedron. If the shales were completely fused and considered as magmas, with the loss of most of their volatiles, their D.I.'s would range from 47 to 68. They would thus lie closer to the base of the tetrahedron than basalts, and in this respect they are more closely allied to intermediate rocks.

Winkler (1957, 1958) concluded that the melt produced by partial fusion of clays free of limestone is the " ichor ", or granite pore-magma, which is the active agent in the process of migmatization and granitization. Our results show that shales containing calcite begin to melt at the same PT conditions as shales free of lime (Tables 1 and 2 ; see also Winkler and von Platen, 1960). This is to be expected from 'phase relations in the system CaO-CO_2-H_2O (Wyllie and Tuttle, 1960a).

There is now evidence that sediments with a wide range of composi-

tions will be partially melted, in the presence of water vapour, at temperatures which could be reached at no great depth in the orogenic belts of the earth's crust (Wyllie and Tuttle, 1960b). Partial melting can be expected in rocks with compositions varying between granite, shale, and impure limestone. This includes arkoses and feldspathic sandstones.

We conclude that partial melting plays an important part in high-grade regional metamorphism, and that the product of fusion for most rocks is a melt with the composition of a granite or granodiorite. If this is the active agent in migmatization and granitization, as suggested by Winkler, then the terms " ichor ", emanation, granitizing fluid, etc., seem to be unnecessary. Any definition of a magma known to the authors would apply to this interstitial melt. It has been evident for some time that the root of " the granite problem " is buried not only beneath mountain ranges, but also in geologists' choice of words.

Summary

Shales begin to melt in the presence of water vapour under pressure at temperatures only slightly in excess of the beginning of melting of granites, and partial fusion produces a liquid of granodioritic composition. The shales are half-melted at temperatures where basalt is just beginning to melt. It is unlikely that a granite magma could be superheated sufficiently to cause more than 50 per cent fusion of a shale, and the liquid produced would still be granodioritic in composition. If a shale were completely melted, the liquid produced would be closer in composition to an intermediate magma than to a basalt, but characteristic chemical differences would distinguish it from any normal igneous magma. These differences would cause the precipitation of aluminous minerals during crystallization. Partial melting can be expected during high-grade metamorphism of shales.

Acknowledgments

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The Composite Sill of Rudh' a' Chromain, Carsaig, Mull

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Abstract

The sill shows the normal type of spatial relationship found in composite intrusions of the British Tertiary, i.e. basic-acid-basic. The basic rocks are Talaidh type tholeiites while the acid rock is a rhyolite. Xenoliths appear to be concentrated in the upper and lower tholeiites. This fact has suggested to the authors of the Mull Memoir that a process of gravitational differentiation resulted in the formation of the basic and acid magmas. The evidence is inconclusive as to what process operated in magma genesis. The mineralogy of the xenoliths found in the sill is described in detail.

I. Introduction

The Loch Scridain area of Mull contains a number of Tertiary composite sills, of which that of Rudh' a' Chromain is perhaps the most striking. Broadly speaking, the rocks comprising these sills belong to a tholeiite-felsite assemblage. The order of injection of the components of the composite sills of Mull appears to have been from basic to acid, this being the normal order in Tertiary composite intrusions.

II. Field Relations

The composite sill of Rudh' a' Chromain is situated about 1½ miles south-west of Carsaig, Isle of Mull, and forms a low headland. The sill is intrusive into the Carsaig Sandstone of Middle Lias age, of which the general dip is to the west at 6 degrees.

The field relations and petrology of the intrusion have been described by Thomas (1922) and Bailey *et al.* (1924). Subsequent field work has in general confirmed the accuracy of the mapping of the Geological Survey, though a further basic sheet has been discovered, considered to be genetically related to the composite sill and is accordingly described with it in the following account.

Within the sill, five members have been recognized, as follows :—

5. Thin tholeiite sheet, up to 1 foot in thickness.
4. Bostonite, earlier in age than the other members (Buist, 1958).
3. Upper tholeiite, up to 8½ feet in thickness, with a chilled top.
2. Felsite, up to 30 feet in thickness, with " sheath-and-core " structure.
1. Lower tholeiite, up to 4 feet in thickness, with a chilled base.

III. Petrography of the Tholeiite Sheets

(*a*) *Lower Tholeiite.* A specimen (of specific gravity 2·81) collected from the base of the sheet is a tholeiite of Talaidh type, composed of laths of labradorite (An_{60}) of average length 0·07 mm., together with augite in laths or grains, scattered patches of calcite and magnetite.

The optical properties of augite are :—$\alpha = 1{\cdot}696$, $\beta = 1{\cdot}700$, $\gamma = 1{\cdot}724$, $\gamma \wedge c = 44°$, $(+)$ $2V = 40°$; composition Ca_{33} Mg_{39} Fe_{28}, according to Hess (1949). An analysis of a rock of this type from the sill is quoted in Bailey *et al.* (1924, p. 17) and repeated in Table 1 of the present account.

A specimen (of specific gravity 2·73) from the middle of the lower tholeiite contains feldspar phenocrysts, skeletal pseudomorphs in chlorite after pyroxene and occasional small vesicles composed of chlorite, calcite, feldspar, a few specks of iron ore and very occasional quartz grains. The rock shows a good deal of alteration consequent on acidification by the later intruded felsite.

(*b*) *Upper Tholeiite.* In its central portions, this sheet is essentially similar to the lower tholeiite, but its margins show evidence of contamination with acid material ; the effects are described below.

A specimen from the lower part of the upper tholeiite shows an intimate mixture of tholeiitic and felsitic material. The former is essentially similar to the typical tholeiite of the sill, being composed of randomly orientated laths of fresh, unfissured labradorite (An_{52}), occasional fresh colourless augite, scattered magnetite and interstitial areas of light brown, finely-divided iron ore. Augite shows alteration to light yellow or yellow-brown " wispy " chlorite and only a little fresh material is found. The felsitic portion of this admixture is composed of scattered, randomly orientated feldspar laths and acicular, altered clinopyroxene in a diffuse matrix of calcite, finely-divided iron ore, and grains of quartz. Occasional vesicles of glass with a partial rim of chlorite are present.

A specimen from the compact surface-layer of the upper tholeiite is a devitrified pitchstone, of specific gravity 2·60, with rounded vesicles filled with quartz and calcite. The thin section shows feldspar as phenocrysts and in the groundmass, together with calcite, iron ore, chlorite, and quartz grains of irregular shape. Determination of composition is not possible, owing to extreme alteration, chiefly to calcite. Chlorite, brownish iron ore and magnetite are found as inclusions within the phenocrysts. The feldspars of the groundmass are normally simply twinned.

(*c*) *Thin Tholeiite Sheet.* The rock is a tholeiite allied to the Talaidh type, but interstitial glass is not found. A specimen (of specific gravity 2·84) from the upper margin of this sheet is composed of bytownite (An_{76}) phenocrysts in a matrix of randomly orientated laths of andesine-labradorite (An_{50}), subordinate augite, chloritic alteration products, and magnetite. Glomero-porphyritic aggregates of bytownite are noted. Augite is colourless or faint brown and occurs in the form of small irregular grains, plates or laths. Its optical properties are :—$\alpha = 1{\cdot}690$, $\beta = 1{\cdot}698$, $\gamma = 1{\cdot}716$, $\gamma \wedge c = 41°$, $(+)$ $2V = 50°$;

composition Ca_{40} Mg_{37} Fe_{23}, according to Hess (1949). A tendency to a cervicorn development of the mineral is noted, similar to that described by Thomas in Bailey *et al.* (1924). Greenish chlorite is abundant in certain areas and also occurs with calcite in irregularly-shaped vesicles.

IV. Petrography of the Felsite

A specimen from the centre of the sheet has a specific gravity of 2·58 and is light grey in colour. The rock is a rhyolite composed of feldspar in phenocrysts and laths, chlorite, iron ore, sporadic quartz, and biotite. The feldspar phenocrysts show Carlsbad twinning. Alteration to calcite and kaolinitic material is characteristic. The feldspars of the groundmass contain inclusions of chlorite, iron ore, and quartz. Greenish chlorite pseudomorphs up to 1·1 mm. in length are present and normally have a rim or partial rim of magnetite. An analysis of this rock is given in Table 1.

The felsite is vitreous in places. A specimen of this type (of specific gravity 2·53) from the central portion of the felsite is dark purple and has a "patchy" appearance in the hand-specimen. The rock is a partially devitrified porphyritic pitchstone containing sporadic phenocrysts of oligoclase (An_{25}), orthoclase (with $2V = 70°$), and altered pyroxene in a groundmass of feldspar, quartz, altered pyroxene, chlorite, and brown glass. Small irregularly-shaped vesicles composed of quartz and calcite are present. Former pyroxene is now represented by calcitic pseudomorphs, which are generally found in the neighbourhood of the oligoclase phenocrysts. The groundmass feldspars show only simple twinning and their composition cannot be accurately determined. Very narrow feldspars with a "sheaf-like" habit are noted.

"Sheath-and-core" structure is locally characteristic of the felsite. A specimen from the centre of a core is the closest approach to the appearance of a typical pitchstone in the sill. The rock is dark brown in colour and has a slightly vitreous lustre due to the presence of scattered irregular yellow-brown glassy areas. The specific gravity is 2·50. The thin section shows sporadic phenocrysts of oligoclase (An_{30}) and pigeonite set in a groundmass of brown glass containing feldspathic aggregates, pigeonite, and rounded grains of black iron ore. Phenocrysts of pigeonite have the form of colourless irregular crystals, sometimes with pointed terminations. The mineral also occurs as acicular crystals. The optical properties are as follows :—$\alpha = 1·709$, $\beta = 1·712$, $\gamma = 1·739$, $\gamma \wedge c = 40°$, $(+)$ $2V = 15°$; composition Ca_{16} Mg_{36} Fe_{48}, according to Hess (1949).

The association of phenocrysts of plagioclase and pigeonite, together with relatively large grains of magnetite in aggregates in the rock is

characteristic. Feldspathic aggregates showing a radiating disposition are also present; these may have a nucleus of pyroxene or glass. An analysis of the rock type is given in Table 1.

The contact of a core with its sheath is composed of phenocrysts of oligoclase (An_{30}), radiating aggregates of altered feldspar, sporadic acicular clinopyroxene, in part chloritized and scattered randomly

TABLE 1.—*Chemical analyses and norms, composite sill of Rudh' a' Chromain, Mull.*

	1.	*2.*	*3.*
SiO_2	71·08	69·66	53·97
TiO_2	0·58	0·56	1·24
Al_2O_3	12·69	12·48	14·65
Fe_2O_3	0·36	1·04	3·62
FeO	4·28	3·56	6·32
MnO	0·20	0·10	0·31
MgO	0·77	0·55	4·49
CaO	1·32	2·48	7·98
Na_2O	2·72	3·40	2·54
K_2O	3·73	2·82	1·52
H_2O+	1·52	2·50	0·94
H_2O-	0·58	1·05	1·92
P_2O_5	0·12	0·07	0·27
	99·95	100·27	100·40
Sp. Gr.	2·58	2·50	2·81

Norms.

	1.	*2.*	*3.*
Q	33·84	31·26	11·04
or	22·25	16·67	8·90
ab	23·05	28·82	20·96
an	6·67	10·28	24·46
C	1·72	—	—
di wo	—	0·81	8·73
di en	—	0·30	
di fs	—	0·66	
hy en	1·90	1·20	13·88
hy fs	6·60	3·96	
mt	0·76	1·62	5·34
il	1·06	1·06	2·28
ap	—	—	0·67
calcite	—	—	1·10
water	2·10	3·55	2·86
Totals	99·95	100·19	100·22

1. Centre of felsite. Analyst: W. H. Herdsman (new analysis).
2. Devitrified pitchstone from core in felsite. Analyst: W. H. Herdsman (new analysis).
3. Tholeiite of Talaidh type, basic margin of sill. Analyst: E. G. Radley (Bailey *et al.*, 1924). Analysis includes $FeS_2 = 0·09$ per cent, $BaO = 0·04$ per cent, and Li_2O a trace.

orientated feldspar microlites set in a brownish devitrified base containing minute dark iron-ore granules and larger flakes of hematite. Small vesicles of irregular shape containing quartz and calcite are present. Determination of the composition of the groundmass feldspars is not possible owing to their largely altered nature.

The sheath material (of specific gravity 2·63) surrounding the cores has a markedly weathered appearance in hand-specimen and is olive-brown in colour. The rock is an acidic or felsitic inninmorite, containing phenocrysts of feldspar and altered clinopyroxene in a devitrified base composed of feldspar, serpentine, magnetite, and chlorite. Clinopyroxene is typically associated with feldspar phenocrysts in aggregates. Under high power, altered clinopyroxene has a minutely " dotted " appearance and shadowy or undulose extinction is noted, while alteration to serpentine along cracks commonly occurs. The feldspars of the groundmass are either randomly orientated or occur in radiating aggregates. Simple twinning is usual. Darkish green bowlingite is a prominent constituent of the groundmass and often shows " mesh " structure. Sporadic areas in the groundmass are composed of feldspar laths in a greenish glass.

V. Xenoliths

Xenoliths found in the various members of the sill can be termed either " cognate " or " accidental " (Thomas, 1922). Xenoliths of the first type, which are dark and coarse-grained, are found in the lower and upper tholeiites. The contact with the enclosing tholeiite is normally rather irregular. These xenoliths are composed of hypersthene and basic plagioclase. Hypersthene occurs in the form of large irregular plates or as somewhat rounded grains. Cleavage is not usually developed and, if present, is discontinuous. Prominent irregular cracks are noted. Narrow irregular veins of chlorite occur within the mineral. The optical properties of the hypersthene are as follows :—α = faint pink = 1·698, β = pale yellow to colourless = 1·707, γ = pale green = 1·712. Plagioclase (An_{62}) occurs in the form of large laths or plates and is occasionally zoned.

The second type of xenolith, namely " accidental ", is present in greatest number in the lower and upper tholeiites. Xenoliths of this type are mainly altered mudstone or shale. These accidental xenoliths can be grouped under the following headings :—(*a*) Aluminous xenoliths, (*b*) Siliceous xenoliths.

(*a*) *Aluminous xenoliths.* In the Rudh' a' Chromain sill, these xenoliths can be divided into two types, namely mullite-buchites and a basic plagioclase-sapphire-spinel assemblage. The mullite-buchites (of average specific gravity 2·67) are dark purplish in hand-specimen and have the appearance of a compacted mudstone. This type of

xenolith is composed of a felted mass of needles and sporadic laths of mullite up to 0·4 mm. in length, together with scattered grains of ilmenite and siderite, set in a colourless or lilac glass of refractive index 1·510. Small spherical vesicles composed mainly of brownish glass and irregular cavities filled with chalcedony are present. George (1924) has investigated the relationship between specific gravity, refractive index and chemical composition of naturally occurring glasses. Using George's curves, it is found that the glass of the mullite-buchite from Rudh' a' Chromain has the following partial chemical composition : SiO_2 67·5 per cent, FeO + Fe_2O_3 4·2 per cent, MgO 1·0 per cent, CaO 2·5 per cent, K_2O 3·9 per cent, H_2O 2·4 per cent. The specific gravity of the glass is 2·42. The magnetic portion of the heavy fraction separated from the mullite-buchite is composed almost entirely of ilmenite, present as sporadic, relatively large blackish grains with a marked steely-blue lustre in reflected light.

The optical properties and specific gravity of mullite from Rudh' a' Chromain are as follows :—α = colourless = 1·649, β = colourless = 1·652, γ = faint pink = 1·662, 2V = 51°, S.G. = 2·969. A chemical analysis, with other analyses for comparison, is given in Table 2.

In other xenoliths of this type, mullite is accompanied by sapphire. The optical properties and a partial spectrographic analysis of the latter are as follows :—N_o = 1·768, N_e = 1·760, 2V = 0° in general, but may be exceptionally up to 8° ; Cr 5,000, V 400, Ni 50, Co 5, Cu 20, Ba 50 (all p.p.m. and ± 50 per cent).

The second type of aluminous xenolith occurs as the modified outer portion of a mullite-buchite. The constituent minerals are anorthite (containing mullite needles), sapphire or spinel or both these minerals.

A thin section of the distinct, though markedly irregular contact between an aluminous xenolith and the upper tholeiite shows the association of basic plagioclase and spinel. Sapphire does not appear in this peripheral zone in the form of discrete crystals and is found only as infrequent inclusions (along with spinel) in plagioclase. Spinel here is purplish-red to almost black in colour. Further in towards the centre of the xenolith, the complete assemblage is seen ; here, spinel is dark green in colour. The earliest formed basic plagioclase (An_{88}) in the outer zones of these xenoliths contains occasional grains of ilmenite, but the most characteristic feature is the large number of enclosed mullite needles. The area of these needles may embrace the whole of a plagioclase crystal, or a narrow peripheral zone may be left free of them. Twinning in the feldspar is generally poorly developed. The next-formed feldspar is less basic than the first and has the form of small irregular crystals, generally moulded on the earlier feldspar. These small feldspars are free from mullite needles and are of

labradorite-bytownite (An_{68}). The last formed feldspar of all is generally skeletal in form and has a composition ranging from acid labradorite to oligoclase.

Spinel shows colour variation. The usual type of spinel noted in the outer zone of these aluminous xenoliths is dark green magnesian hercynite. An analysis of this mineral, quoted by Bailey *et al.* (1924),

TABLE 2.—*Analyses of mullite and spinel, Rudh' a' Chromain, Mull.*

	1.	*2.*	*3.*	*4.*	*5.*
SiO_2	28·65	29·04	29·36	0·77	1·07
TiO_2	0·51	0·79	1·12	0·50	0·10
Al_2O_3	70·10	69·63	69·05	60·84	61·95
Fe_2O_3	0·32	0·50	0·86	4·26	5·05
FeO	—	—	—	24·00	15·04
MnO	—	—	—	0·15	—
MgO	0·05	—	0·10	9·37	16·22
CaO	—	—	—	0·36	0·20
Na_2O	0·12	0·18	—	—	—
K_2O	0·03	0·06	—	—	—
H_2O	—	—	—	0·14	0·18
	99·78	100·20	100·49	100·39	99·81
Sp. Gr.	2·969	—	—	3·837	4·050

*Analyses, recalculated as atomic proportions.**

	1.	*2.*	*3.*	*4.*	*5.*
Si	2·043	2·063	2·079	0·021	0·028
Al	5·886	5·821	5·755	1·921	1·887
Fe'''	0·017	0·026	0·051	0·087	0·099
Fe''	—	—	—	0·536	0·325
Mg	—	—	0·013	0·376	0·631
Ca	—	—	—	0·011	0·006
Na	0·017	0·026	—	—	—
K	—	0·009	—	—	—
H	—	—	—	—	—
Ti	0·026	0·043	0·060	0·010	0·002
Mn	—	—	—	0·003	—

1. Mullite separated from xenolith of mullite-buchite in lower tholeiite, composite sill of Rudh' a' Chromain. Analyst: D. S. Buist (new analysis).
2. Mullite from Seabank Villa, Mull (contained 0·51 per cent corundum and 0·1 per cent rutile, which were determined separately). Analyst: E. G. Zies (Bowen, Greig, and Zies, 1924).
3. Mullite from the crowns of glass pots. Analysts: N. L. Bowen and M. Aurousseau (Bowen, Greig, and Zies, 1924).
4. Dark green magnesian hercynite from contaminated aluminous xenolith, Nun's Pass, Carsaig, Mull. Analyst: E. G. Radley (Bailey *et al.*, 1924).
5. Purplish to dark blackish-brown magnesian hercynite from contact of aluminous xenolith and lower tholeiite, Rudh' a' Chromain, Mull. Analyst: D. S. Buist (new analysis).

* On basis of thirteen oxygens for mullite and four oxygens for spinel.

recalculated to atomic proportions, is given in Table 2. A darker coloured variety of spinel is found near the periphery of these aluminous xenoliths ; the mineral varies in colour from light or dark purplish-red to blackish-brown. The specific gravity is 4·05, while the refractive index is 1·784. X-ray powder data show that the cell dimensions measure 8·11 Å. An analysis of a dark blackish-brown spinel from the contact of an aluminous xenolith and the lower tholeiite is given in Table 2.

(*b*) *Siliceous xenoliths.* Little need be added to Thomas' description (Bailey *et al.*, 1924) of the sandstone xenoliths, but three examples will be described in detail ; the last two are of somewhat abnormal type.

The first example is of fossiliferous shaly sandstone of specific gravity 2·71. Quartz is a prominent constituent in angular to sub-angular grains and in the interstices between these, dark brown shaly material occurs, composed of calcite (in irregularly shaped patches or in a form suggestive of a pseudomorph after feldspar), white mica and medium to dark reddish-brown pleochroic biotite.

The second example is a lightish grey-white xenolith (of specific gravity 2·53), taken from the upper tholeiite. The rock is a medium-grained sandstone with a somewhat rough and altered appearance in hand-specimen, but differing little in its essentials from the Carsaig Sandstone into which the sill is intruded. The polished surface shows a number of brownish elongate glassy areas with a pronounced flow-structure. These glassy areas probably represent invasions of material derived from the upper tholeiite.

The third example of a sandstone xenolith is from the lower margin of the upper tholeiite sheet. In hand-specimen, the rock is compact and medium grey in colour, with a greasy lustre suggestive of a quartzite ; the specific gravity is 2·63. Microscopic examination shows that an original feldspathic rock has been replaced, to a marked degree, by quartz. In ordinary light, the rock appears to be composed of randomly orientated feldspar laths of average length 0·35 mm., set in a dark brown ferruginous and chloritic base, which contains sporadic zircon. Between crossed-nicols, however, the rock is composed predominantly of interlocking grains of " strained " quartz, together with untwinned plagioclase. This is an example of metasomatic alteration. The outlines of the original feldspar are generally preserved. Vesicles, composed of a central core of calcite or quartz, surrounded by " wispy " darkish green chlorite, are present in the rock.

VI. Petrogenesis and Mechanism of Emplacement

To account for the origin of the components of composite sills and other intrusions in south-west Mull which contain " cognate " and " accidental " xenoliths, the authors of the Mull Memoir (Bailey *et al.*,

1924) invoke the hypothesis of gravitational differentiation, as follows :—

> " Bearing in mind the non-metamorphic character of the sediments which furnished many of the characteristic xenoliths, we are justified in placing some boundary of the magmatic reservoir in the space interval between Tertiary lavas and the pre-Mesozoic floor of gneiss and granite. It is suggested that the roof of the reservoir may have roughly coincided with the base of the lavas in part of the subsidence area of Central Mull, east of the head of Loch Scridain. It seems that the reservoir was of considerable depth, since it was able, in many instances, to supply material of widely differing composition to individual composite sills. A concentration of xenoliths, both " cognate " and " accidental ", in the more basic portion of the magma is exhibited in the Rudh' a' Chromain sill and certain other good examples. Such a concentration is in keeping with the idea of gravitational differentiation within a magma-basin of the type here advocated."

This concentration of xenoliths in the basic rock of the Rudh' a' Chromain sill (and others) is not of necessity evidence favouring gravitational differentiation and might merely be an indication of the greater facility of the basic magma (because of its considerably higher temperature compared with that of the acid magma) to attack the walls of the reservoir and incorporate the fragments of the country rocks as xenoliths.

The emplacement of the sill was guided by the bostonite sheet—the latter probably contemporaneous with the " Central " basalts, as suggested by Bailey *et al.* (1924). The first intrusive act resulted in the emplacement of a basic sheet approximately 10 feet in thickness (the potential upper and lower tholeiites) and the thin basic sheet, in flanking positions with regard to the bostonite. After an interval, probably short, during which the basic material had cooled to a certain degree, the felsitic member was intruded along a " plane of weakness " in the centre of the combined upper and lower tholeiites.

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Rolled Garnets from Morar, Inverness-shire

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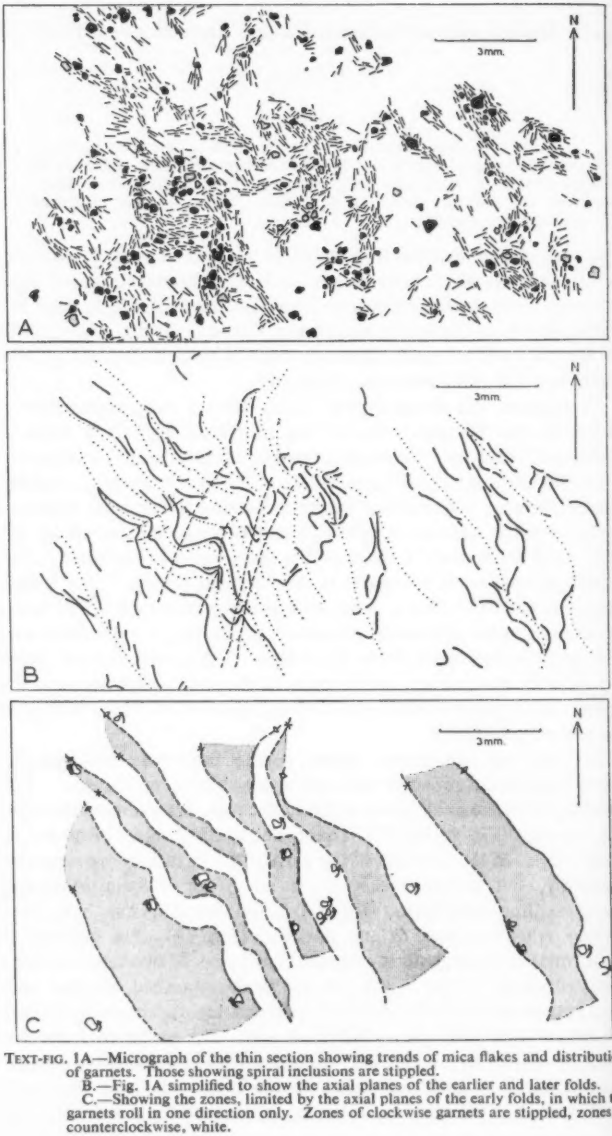
Abstract

A thin section of garnetiferous pelitic schist is described in which the rotation of the garnets is in either of two directions. Microfolds are seen in the slide and the garnets were rotated in opposite directions on opposite limbs of the folds. Their relation to minor structures is suggested.

The rock from which this thin section is taken belongs to the sequence of Moine rocks which outcrops on the Scottish mainland east of Skye. It comes from the Garnetiferous Pelitic Subgroup of the Striped and Pelitic Group designated by Richey and Kennedy (1939). The rocks lie in Barrow's garnet grade although there is some retrogressive metamorphism with the formation of chlorite.

Throughout the Striped and Pelitic Group two generations of structures can be seen : to the first generation belongs a series of elongated folds which are accompanied by an axial plane cleavage and a lineation or quartz rodding parallel to the axes. At certain localities, recrystallization in the pelitic members parallel to the axial planes of these folds has been so complete that it becomes difficult to see the folds, and from such a locality this specimen was collected. The second generation of structures is the " herring-boning " that Richey described in 1948. This is sometimes accompanied by folds of larger dimensions in the competent psammitic bands, but is more often seen only as little chevron folds in the pelites. The specimen from which the slide is taken shows quartz rods of the first generation and the schistosity is crinkled along the axial planes of folds which belong to the second generation.

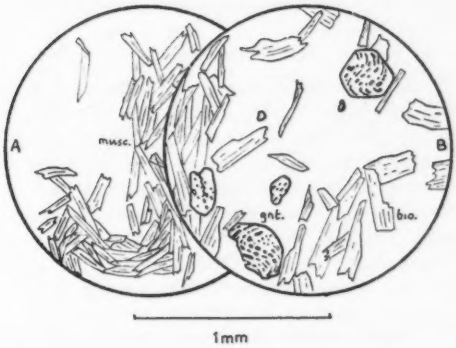
The rock contains garnet, quartz, biotite, muscovite, sodic plagioclase, clinozoisite, opaque ore and a few blades of chlorite. The garnets, which are small, rarely reaching 0·5 mm., are scattered throughout the slide (Fig. 1A) and invariably contain inclusions of quartz or opaque ore. In the majority of the garnets the inclusions are arranged randomly, but in some they show an *S*- or *Z*-form indicating paracrystalline deformation (Fig. 1A) ; individual crystals have been rotated either clockwise or anti-clockwise (Fig. 2B). The orientation of the micaceous minerals is irregular, but traces of two sets of microfolds, which are shown in Fig. 1B, can be distinguished, the first with axial planes trending N.W.-S.E. with reference to an arbitrary North and the second with axial planes N.N.E.-S.S.W. The earlier folds are the better developed set, and are elongated and nearly isoclinal in style, with sharp crests and a small angle between the limbs. The later set have a gentle wave-like form and bend the axial planes of the first folds. The



TEXT-FIG. 1A—Micrograph of the thin section showing trends of mica flakes and distribution of garnets. Those showing spiral inclusions are stippled.
B.—Fig. 1A simplified to show the axial planes of the earlier and later folds.
C.—Showing the zones, limited by the axial planes of the early folds, in which the garnets roll in one direction only. Zones of clockwise garnets are stippled, zones of counterclockwise, white.

micas are unstrained and form polygonal outlines to the fold crests (Fig. 2A) ; hence it appears that the mimetic crystallization observed by Richey (1948) took place subsequent to these two " phases " of folding.

The thin section can be divided into zones separated by the axial planes of the earlier folds (Fig. 1C). It is found that in any one zone all the garnets whose sense of movement can be determined have been



TEXT-FIG. 2A.—Microfolds of the later phase of folding showing the effect of mimetic crystallization.

B.—Garnets showing trains of inclusions, those in the upper garnet arranged in Z-form, those in the lower, in S-form.

rotated in the same sense, but that in the adjoining zones the sense of rotation is the reverse. In short, the garnets have been rotated in opposite directions on either flank of each fold, the sense of movement in each zone being towards the fold-hinge. In this way one can, from the evidence presented by the spiral fabrics in the garnets of adjacent zones, designate the axial planes which separate these zones as belonging to " antiforms " or " synforms ", and this is diagrammatically shown in Fig. 3. The growth of garnet is thus related to the formation of the early folds, whilst during the later phase the mineral was not growing although it was stable.

The minor folds seen in the field are almost isoclinal and, in these pelites, further recrystallization parallel to the axial planes would produce a texture parallel to and indistinguishable from the effect of mimetic crystallization on the primary mineral orientation. In other words, but for the evidence of the rolled garnets it would be difficult in the section to tell whether the mineral layering were recrystallized

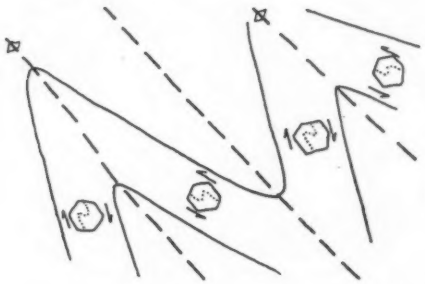
bedding or the limbs of isoclinal folds ; their presence, however, enables the following history to be deduced from the slide :—

(1) Formation of narrow, almost isoclinal folds ; paratectonic growth of garnet.

(2) Bending of first folds by the second ; no growth of garnet.

(3) Mimetic crystallization.

Presumably the growth of these garnets can be linked with the



TEXT-FIG. 3.—Diagram to show the sense of movement of the garnets in relation to the axial planes of the earlier folds

movements which produced the minor folds of isoclinal type, whilst the bending corresponds to Richey's " herring-boning ".

Elsewhere there is evidence that these are not the earliest stages in the mineralogical sequence, for these relatively small garnets which grew syn-kinematically are preceded by a generation of larger garnets which enclose the remains of an even earlier schistosity.

The author is indebted to Professor J. Sutton and Dr. Gilbert Wilson for help with this paper.

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Some Erratics from Teifi Estuary, Western Cardiganshire [1]

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Abstract

A petrographical description is given of some erratics from the Teifi Estuary, Cardiganshire. Rock types present include granite from Ireland ; greisen and Eskdale granite from the Lake District ; granophyre, picrite, diabase and gneiss from the Lleyn Peninsula ; granophyre and diabase from Merionethshire.

Introduction

Erratics have been recorded in the past from various localities along the coast of Wales (Keeping, 1878, 1882 ; Jehu, 1904, 1909 ; K. E. Williams, 1927). In 1957, Professor O. T. Jones recognized that among erratics occurring on the north side of the Teifi Estuary, half a mile south of Gwbert Hotel, Cardiganshire, were some unusual types. He suggested that the writer should make a further collection from the same locality. The erratics collected by Professor Jones and the writer include the igneous and metamorphic rocks described below. Although the petrological matching of an erratic with a rock found *in situ* is not proof of their identity, it is probable that the true sources of some or all of the present erratics have been found.

Carboniferous Limestone specimens collected contain fossils which have been identified by Dr. W. H. C. Ramsbottom as *Lithostrotion martini* and a kind of Athyrid, but their provenance is uncertain. Some specimens of hard calcareous sandstone contain lamellibranch shells, identified by Mr. R. V. Melville and Dr. R. Casey as *Neomiodon* and *Staffinella*, genera which occur in association in the Great Estuarine Series of the Hebrides.

Petrography of the Erratics

Granites.—A medium-grained cream coloured granite (E 27699) * contains large crystals of microcline enclosing albite usually somewhat altered to kaolin and sericite, a small amount of microperthite, quartz in granular aggregates, numerous plates of reddish-brown biotite enclosing zircon surrounded by pleochroic haloes, and ragged flakes of muscovite in parallel intergrowth with the biotite. Some of the biotite crystals are quite fresh and others have been pseudomorphed by chlorite. Magnetite and prisms of apatite are accessory minerals. In colour and texture the rock resembles a muscovite-biotite-granite

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* Numbers in parenthesis refer to sliced rocks in the Geological Survey collection.

(I 1233) from Dun Loaghaire, near Dublin, Ireland. Granite boulders possibly derived from the east of Ireland have been recorded by Howard and Small (1899) from Skomer Island, Pembrokeshire.

A moderately coarse-grained pink granite (E 28824) contains quartz, a large amount of perthite and some albite-oligoclase, large crystals of biotite which have been completely chloritized, small flakes of muscovite and irregular crystals of iron-ore. Accessory minerals

TEXT-FIG. 1.—Sketch map showing localities mentioned in text.

include apatite in needles and prisms and a little sphene. Specimens of Eskdale granite (E 16856–7) match the erratic.

Greisen.—A medium-grained creamy-buff rock (E 27704) consists essentially of a mosaic of quartz crystals and muscovite in large tablets and aggregates of smaller flakes. It is very similar to the greisen (E 15745) of Grainsgill Beck, Cumberland, described by Harker (1895) who mentions the occurrence of a similar greisen associated with the Foxdale Granite, Isle of Man.

Granophyres.—A buff coloured rock (E 28821) contains crystals of oligoclase, somewhat sericitized, in a groundmass of quartz, micropegmatite and small stumpy prisms of albite. Clusters of flakes of brownish-green biotite and fibrous amphibole also occur, with patches of iron-ore and prisms of apatite as accessory minerals. The rock matches specimens of porphyritic granophyre (E 15644, 15920, 17426), from Llanbedrog, Caernarvonshire.

A greenish-grey medium-grained granophyre (E 28820) contains small idiomorphic crystals of albite and orthoclase set in a groundmass of microgranular quartz and small laths of albite surrounded by micrographic intergrowths of quartz and feldspar which are sometimes microspherulitic. There are some patches of calcite, and chlorite is abundant in granules and large aggregates. Accessory minerals include small crystals of zircon, a few crystals of iron ore and tiny needles of apatite. Specimens of Crogenen granophyre (E 28837, 28861) from the Arthog-Dolgelly area match the erratic. These rocks have been described by Cox and Wells (1920, p. 290 ; 1927, p. 297).

Picrite.—A medium-grained lustre-mottled greenish-black picrite (E 27700) contains rounded grains of olivine in various stages of alteration to serpentine, large ophitic plates and fibrous hornblende, a small amount of colourless augite in irregular plates, flakes of brown biotite, and granules of iron ore. The feldspar has been completely saussuritized. The rock matches a hornblende-picrite (thin section : Lindsay Collection, 1239) from the west side of Mynydd Penarfynydd, Caernarvonshire. Two previous discoveries of hornblende-picrite have been made in the neighbourhood. Hicks found a boulder at Porthlisky, near St. David's, Pembrokeshire, which Bonney (1885, p. 518) believed to be derived from Anglesey or the Lleyn Peninsula. A boulder of picrite similar to the Penarfynydd specimens was collected by Jehu (1904, p. 79) on Strumble Head, Pembrokeshire.

Diabases.—A greyish-green diabase with darker patches (E 27701) contains some serpentinous pseudomorphs which may have been original olivine. The most abundant minerals are brown and green hornblende in large plates and fibres and colourless to very pale brownish-green augite often showing crystal boundaries. There is some comparatively fresh albite, which has formed late and encloses augite and hornblende crystals. A few skeletal crystals of ilmenite also occur. Harker (1889, p. 94) has described a hornblende-diabase (thin section : Sedgwick Museum Collection, No. 718), from near Ffynnoncefn, a quarter of a mile north of Rhiw, Caernarvonshire, which is very similar to this erratic.

A medium-grained diabase (E 27703) contains ragged prisms and irregular patches of albitic feldspar. Original ferromagnesian minerals have been completely altered to granules of carbonate and chlorite. Numerous small crystals of iron ore and large areas of interstitial chlorite and quartz also occur. The rock closely resembles an albite-diabase (E 6985) from Traeth-Llan, near Aberdaron, Caernarvonshire.

Laths of albitic feldspar are the main constituent of a fairly fine-grained altered quartz-diabase (E 27702). Interstitial material consists of irregular grains of quartz, green fibrous hornblende, patches of chlorite and scattered crystals of iron ore. Similar specimens of

quartz- diabase are described by Cox and Wells (1920, p. 280 ; 1927, p. 299) from the Arthog-Dolgelly district.

Dark patches in a medium-grained greenish-grey diabase (E 28822) are seen under the microscope as large masses and fibres of pale bluish-green secondary hornblende and small crystals of pale brown augite. Large twinned laths of andesine feldspar are altered to saussurite. Granular iron ore is fairly abundant in patches and clusters of stumpy crystals of epidote also occur. The rock matches a diabase (E 16636) from Mynydd Gader, near Dolgelly, Merionethshire.

Gneisses.—A greenish-grey foliated rock (E 28827) contains a fine-grained mosaic of quartz and albitic feldspar with some coarser-grained patches, and folia of small flakes of muscovite and chlorite, and grains of iron ore. A similar type of foliated rock (E 28830) is cut by veins of quartz and pale-green chloritic material.

These rocks are similar to sheared gneiss (E 18353–4) from the Pre-Cambrian Mona Complex at Dwrhos near Aberdaron (Matley, 1928).

The writer wishes to thank Professor O. T. Jones for his helpful advice and for initiating the study and Professor C. E. Tilley and Dr. S. O. Agrell for the loan of thin sections.

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Geological Survey and Museum, London.

CORRESPONDENCE

AGE OF METAMORPHISM OF MOINE SCHISTS

SIR,—Measurements of geological age have been carried out on a number of pelitic Moine schists and gneisses using the potassium-argon method. These rocks, though quite similar mineralogically, from widely separated localities throughout the Highlands of Scotland, were made available from the collection of the Geological Survey of Great Britain (Edinburgh Office). All yielded samples of fresh mica, mainly biotite, formed during the metamorphism. Measurements were made on these micas. Details of the localities are given in Table 1.

TABLE 1

Geol. Surv. No.	Map		County	Locality
	1 in.	6 in.		
U 1368	43	177 NE.	Argyll	400 feet SW. of where river leaves Loch Assapol, Ross of Mull.
R 83	52	25 SE.	Argyll	Near H.W.M. on east side of small bay 990 yards S. 37° W. of Glenmore Bridge, Ardnamurchan.
S 1716	52	157 NE.	Inverness	NE. corner of Lochan Bad na Sgitheiche 850 yards E. 28° S. of △52 Ard Molach, Moidart.
U 2228	71	77 SE.	Inverness	480 yards SSE. of Eilean A' Mhiuneil, Loch Hourn.
V 261	82	84 NE.	Ross	Burn a few yards above wood ¾ mile NE. of Achnasheen.
V 1127	109	46 NW.	Sutherland	⅞ mile N. of boathouse Loch Coire nam Mang, 6¾ miles NW. of Kinbrace.
V 1398	74	47 NW.	Inverness	On hill slope ⅝ mile ENE. of Coignafeuinternich, River Findhorn 8½ miles above Tomatin.

Argon was extracted from the micas by fusing them without any flux, in a molybdenum crucible. The mixture of evolved gases which included the argon was purified using red hot titanium sponge, which has the property of being able to absorb large quantities of certain gases and retain them on cooling to room temperature. Argon is not taken up in this way and was further cleansed by absorbing it on to active charcoal, then pumping on this for about half an hour to remove traces of hydrogen. After repeated cleanings the total volume of argon was measured using a McLeod gauge.

Potassium analyses were carried out using the flame photometer, each sample being subjected to six separate potassium determinations. Each determination involved thirty flame photometer readings, the figure for each determination was obtained from the average of all thirty photometer readings, and the values quoted in the following table are in each case the average of all six done in this manner. K_2SO_4 standards were used to calibrate the photometer.

Using the mass spectrometer, the A^{40}/A^{36} ratios of argon extracted from the micas were determined, and used to find the percentage contamination of the sample by atmospheric argon. This was done by comparing the isotopic ratio of the sample with that of argon obtained from the atmosphere.

The two constants used in the calculations are λ_ε the electron decay

constant, here assumed to be $0{\cdot}585 \times 10^{-10}$ yr.$^{-1}$ and $\lambda\beta$ the β^- emission decay constant $4{\cdot}76 \times 10^{-10}$ yr.$^{-1}$.

The results are summarized in Table 2.

TABLE 2

Specimen No.	Analysis Ref. No.	Wt. of sample loaded (gm.)	K_2O %	Total vol. of argon (mm.^{3})	Atmospheric contamination %	Apparent age, million years
U 1368	KA/30	6·0013	8·32	0·847	7	429
R 83	KA/48	8·5126	8·67	1·220	4	434
S 1716	KA/45	7·2330	7·99	0·990	7	431
U 2228	KA/2	6·0544	6·82	0·668	2	434
V 261	KA/33	7·6398	7·36	1·032	14	426
V 1127	KA/31	8·1040	8·50	1·20	7	441
V 1398	KA/25	5·1634	7·55	0·632	2	434

Average 433 $\pm$ 8 (range of extremes).

Errors in these apparent ages introduced by the experimental procedure are believed not to be more than $\pm$ 5 per cent.

These results indicate that the Moine Series of Scotland was subjected to regional metamorphism during Silurian (Holmes, 1959) times. Their consistency suggests that they have not since then undergone any disturbances, apart from localized intrusions, that have resulted in the serious loss of radiogenic argon from the micas formed as a result of this metamorphism. The existence of any previous episodes of regional metamorphism has not been proved or disproved by this work, neither has information been obtained on the actual age of the Moine sediments themselves.

Three hypotheses regarding the age of metamorphism, based on geological evidence, have been advanced (Phemister, 1948), namely, (*a*) Lewisian, (*b*) post-Lewisian but pre-Torridonian, (*c*) Caledonian. Much of present-day opinion is in favour of the view (*c*), which these results support.

This is part of a project sponsored by the Department of Scientific and Industrial Research. I am indebted to the Royal Dutch Shell Oil Company for a studentship.

JOHN A. MILLER.

DEPARTMENT OF GEODESY AND GEOPHYSICS,
CAMBRIDGE.

29th September, 1960.

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AGE OF MOUNTSORREL GRANITE

SIR,—The granite or granodiorite of Mountsorrel, situated some seven miles to the north of Leicester, England, is a large intrusive boss more than a mile in diameter, and projects through the Trias to give a series of isolated outcrops. Two varieties of this coarsely-crystalline rock exist and are distinguished by their pink and grey feldspars.

The age of this intrusion has long been in doubt. It is older than the

overlying Triassic sandstones, but is not similar to other igneous rocks of Charnwood Forest. It appears to be later than the Charnian syenites since it has not been subjected to N.W.–S.E. movements which affect them. At Kinchley, a diorite said to be associated with the granite is intruded into slates which are probably members of the Swithland group, the uppermost member of the Charnian pyroclastic sequence. On the other hand the granite is itself cut by dolerite dykes of Carboniferous age. Thus it is pre-Carboniferous and younger than some pre-Cambrian rocks of Charnwood Forest. Petrologically it has affinities with the Caledonian group of granites. The above situation was summarized by Watts (1947).

Measurements of geological age have been carried out on three samples of Mountsorrel granite using the potassium-argon method. Biotites were isolated from the material by a process of grinding, sieving, and magnetic separation, then analysed for their potassium oxide content flame-photometrically (see Table I).

TABLE I.—*Results of potassium oxide analyses*

Analyses Reference number	K_2O per cent (each average of thirty readings)	K_2O per cent of sample (average of six determinations)
KA/40	5·53, 5·50, 5·35, 5·35, 5·50, 5·50,	5·46
KA/52	6·18, 6·17, 6·18, 6·18, 6·18, 6·18,	6·18
KA/53	5·65, 5·66, 5·66, 5·66, 5·66, 5·67,	5·66

The extraction of argon, its purification and measurement were carried out in an argon line connected directly to the mass spectrometer. Thus samples were corrected for their atmospheric argon content without any risk of contamination arising during the transfer of gas from the former to the latter. This system also has the advantage that any residual background in the argon line itself may be examined and corrected for. The results together with errors and physical constants used in the calculations are given in Table II.

Results of potassium-argon dating thus indicate that the Mountsorrel Granite is Caledonian in age.

The total error resulting from uncertainties in volume, potassium oxide content, and A^{40}/A^{36} ratio of the sample should not exceed $\pm$ 5 per cent.

TABLE II

Sample	Record Ref.	Vol. of argon (mm.³) / Wt. of sample (gm.)	Atmospheric contamination, per cent	δV_r	δt	Age (t) (millions of years)
KA/40	G/21	0·0839	3·1	0·2	3·63	404
KA/52	G/17	0·0958	4·1	0·3	3·63	403
	G/18	0·101	5·6	0·5	3·74	417
KA/53	G/19	0·0894	5·5	0·5	3·46	405
	G/20	0·088	4·6	0·2	3·62	402

$\lambda\beta = 4{\cdot}72 \times 10^{-10}\text{yr.}^{-1}$ $\lambda\alpha = 0{\cdot}584 \times 10^{-10}\text{yr.}^{-1}$

δV_r = percentage standard deviation in volume of radiogenic argon due to uncertainty in isotopic ratios of specimen.

δt = error in age (in million years) due to a 1 per cent error in proportion of K_2O or volume of radiogenic argon.

We wish to thank the Mountsorrel Granite Company for allowing us to visit their quarry and collect samples. We are indebted to the Department of Scientific and Industrial Research for financing the work and to the Royal Dutch Shell Oil Company for a studentship to one of us (J. A. M.).

JOHN A. MILLER.
JILL S. PODMORE.

DEPARTMENT OF GEODESY AND GEOPHYSICS,
CAMBRIDGE.
1st October, 1960.

REFERENCE

WATTS, W. W. 1947. *Geology of the ancient rocks of Charnwood Forest, Leicestershire*, Leicester.

DISCOVERY OF COPPER MINERALS IN LEWISIAN GNEISS

Sir,—The presence of traces of copper ore has not, so far as the writers are aware, been reported in the Lewisian gneiss of north-west Scotland. An occurrence of copper-bearing minerals was found by us recently in the gneiss about one quarter of a mile south-east of the Bay of Clachtoll, near Stoer, Sutherland (Grid reference : 041267). The minerals form patches in the gneiss some two square inches in diameter over an area of about one square yard. Near the mineral locality there is a talc, anthophyllite dyke with which the occurrence may or may not be associated. Intensive search in the neighbourhood has failed to locate any other occurrence of copper minerals in the Lewisian gneiss.

The presence of copper minerals occurring in the Torridonian beds of this region was reported by Sir Lewis Fermor in the Geological Magazine (*Geol. Mag.* lxxxviii, 1951, pp. 215–218), who gave their locus " on the promontory separating the Bay of Clachtoll on the south from the Bay of Stoer on the north ". It is not suggested, however, that there is necessarily any relationship between these occurrences.

Although we failed to find further traces of copper minerals in the Clachtoll area, minute quantities were found by Mr. A. K. MacLeod of Inverkirkaig, in the gneissic rocks of a quarry alongside the road bordering Loch an Eisg-brachaidh, about a mile south of Loch Kirkaig (Grid reference: 073178). Unfortunately, soon after these specimens were found, the quarry suffered a severe rock-fall, rendering further investigation both unproductive and dangerous. Smears of copper compounds were, however, detected on rock faces in another small quarry at the top of the hill in which the occurrence of harmotome crystals has been recorded by Waterston (*Miner. Mag.*, xxx, 1953, pp. 136–138).

The minerals at Clachtoll consist principally of *brochantite* [$Cu_4(OH)_6SO_4$], accompanied by small amounts of quartz, iron oxide and chlorite ; but *malachite*, *chalcocite*, *covellite* and *bornite* were also found in associated specimens. The samples found at Eisgbrachaidh consist mainly of *malachite* accompanied by *haematite*.

The writers are indebted for these identifications to Dr. Diane C. Knill and Mr. B. R. Young of the Geological Survey, London ; and to Mr. P. McL. D. Duff of the Grant Institute of Geology, Edinburgh, whose help is greatly appreciated.

Rock specimens containing these cupriferous deposits have been presented to the Geological Museum, London (Nos. M.1. 30217/30371/30373) ; and to the Geological Department of the Royal Scottish Museum, Edinburgh (No. 1959. 5).

ALEX J. BOYD AND JAMES M. CRICHTON.

INVERKIRKAIG,
NR. LOCHINVER,
VIA LAIRG,
SUTHERLAND.
15th August, 1960.

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